Volume 7, Number 1
January–February 1999

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

Official English Translation of *Stratigrafiya, Geologicheskaya Korrelyatsiya*

Editor-in-Chief
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http://www.maik.rssi.ru

Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

МАИК "НАУКА/INTERPERIODICA" PUBLISHING

Stratigraphy and Geological Correlation

(Stratigrafiya, Geologicheskaya Korrelyatsiya)

ISSN: 0869-5938

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PUBLISHER
МАИК "НАУКА/INTERPERIODICA"
PROFSOYUZNAYA UL. 90, MOSCOW, 117864 RUSSIA
PHONE: 7 (095) 336-1600; FAX: 7 (095) 336-0666
E-mail: compmg@maik.rssi.ru
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SUBSCRIPTION PRICES (1999, Volume 7, 6 issues)
US $906.00 (North America), $1041.00 (elsewhere). A 5% discount is given on 2-year orders. Subscriptions may be entered at any time. Orders may be placed at agencies or directly. There is a 5% discount to all subscription agencies. Cancellations not allowed except for duplicate payment. Prepayment required. Advance payment required in US$/US bank. MasterCard/Visa accepted. Make checks payable to *Stratigraphy and Geological Correlation* or Interperiodica.

Stratigraphy and Geological Correlation (ISSN: 0869-5938) is published bimonthly for US $906.00 (North America), outside North America for $1041.00 per year by MAIK "Nauka/Interperiodica", 5724 Hwy. 280 East, Birmingham, Alabama 35242-5724 USA. Periodical postage paid at Birmingham, Alabama, and additional mailing offices. POSTMASTER: Send address changes to *Stratigraphy and Geological Correlation*, P.O. Box 1831, Birmingham, Alabama 35201 USA.

The journal was founded in 1993.

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Age Limits of High-Grade Metamorphic Supracrustal Complexes in the Central Aldan Shield: Sm–Nd Isotopic Data

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Received March 18, 1997

Abstract—First investigation results on Sm–Nd isotope systems in clastic metasedimentary and metavolcanic rocks from the high-grade metamorphic supracrustal complexes of central Aldan Shield are used to approximately evaluate their maximum age boundary. It is established that the Upper Archean rocks (model age $T_{Nd}(DM)$ = 3.3–2.6 Ga) served as a principal source material for metapelites of the Kurumkan Group; metapelites and metagraywackes of the Amedichi Group; and also for the metagraywackes of the Kurulta, Zverev, and Seim groups. Provenances with the Early Archean model age ($T_{Nd}(DM)$ = 3.5 Ga) are found to be appropriate only for few rock samples from the Kurumkan and Kurulta groups. The Early Proterozoic age ($T_{Nd}(DM)$ = 2.5–2.2 Ga) is inferred for the protolith material of supracrustal rocks of the Chuga, Kholbolokh, Kyurikan, Idzhek, and Fedorov groups. The data obtained suggest that supracrustal rocks of the Aldan granulite–gneiss and East Olekma granite–greenstone terranes were formed not in the Early Archean, as it was formerly believed, but at the Late Archean and/or Early Proterozoic stages of geologic evolution of the shield.

Key words: *Early Precambrian, high-grade metamorphic complexes, sedimentary rocks, Aldan Shield, age, Nd isotope systematics.*

INTRODUCTION

In recent schemes of tectonic zoning of the Aldan Shield (*Dokembriiskaya ...*, 1988; *Precambrian geology ...*, 1993), the central granulite–gneiss terrane is considered as localized between the western Olekma and eastern Batomga granite–greenstone areas (Fig. 1). The Aldan granulite–gneiss terrane (AGGT) is about 600 × 300 km in area and composed of supracrustal rocks metamorphosed under conditions of the granulite facies (*Rannii dokembrii ...*, 1986). Many researchers consider the high-grade metamorphic rocks of the shield as Early Archean in age (Dzevanovskii, 1958; Cherkasov, 1979; Neelov and Mil'kevich, 1979; Frumkin, 1981; Reutov, 1981; Salop, 1982; *Rannii dokembrii ...*, 1986; *Evolyutsiya ...*, 1987; *Dokembriiskaya ...*, 1988; *Precambrian Geology ...*, 1993; Popov and Smelov, 1996). At the same time, some scientists believe that the rocks were formed in either the Late Archean (Rudnik, 1975) or Early Proterozoic (Tugarinov *et al.*, 1967). Available geochronological data on the AGGT supracrustal complexes are doubtful so far (*Drevneishie ...*, 1989; *Precambrian Geology ...*, 1993), and their chronology is defined either from relations with epochs of endogenic activity, such as events of deformation, metamorphism, and magmatism (e.g., Rudnik, 1975; *Rannii dokembrii ...*, 1986), or on the basis of "metamorphic" and "stratigraphic" criteria. In this work, we consider the age problem of these high-grade metamorphic complexes using the results of Sm–Nd isotopic study.

BRIEF CHARACTERISTICS OF SUPRACRUSTAL COMPLEXES

Various problems of geology and stratigraphy of high-grade metamorphic complexes of the Aldan Shield have already been discussed in many publications (Korzhinsky, 1936, 1939; Dzevanovskii, 1958; Mironyuk *et al.*, 1971; Salop and Travin, 1974; Rudnik, 1975; Cherkasov, 1979; Neelov and Mil'kevich, 1979; Frumkin, 1981; Reutov, 1981; *Rannii dokembrii ...*, 1986; *Dokembriiskaya ...*, 1988; *Precambrian Geology ...*, 1993; and others). In order to present and discuss the Sm–Nd isotopic date in this paper, we use the scheme of consecutive geological events suggested by the scientific team guided by V.L. Dook and V.I. Kitsul (*Rannii dokembrii ...*, 1986; *Dokembriiskaya ...*, 1988;

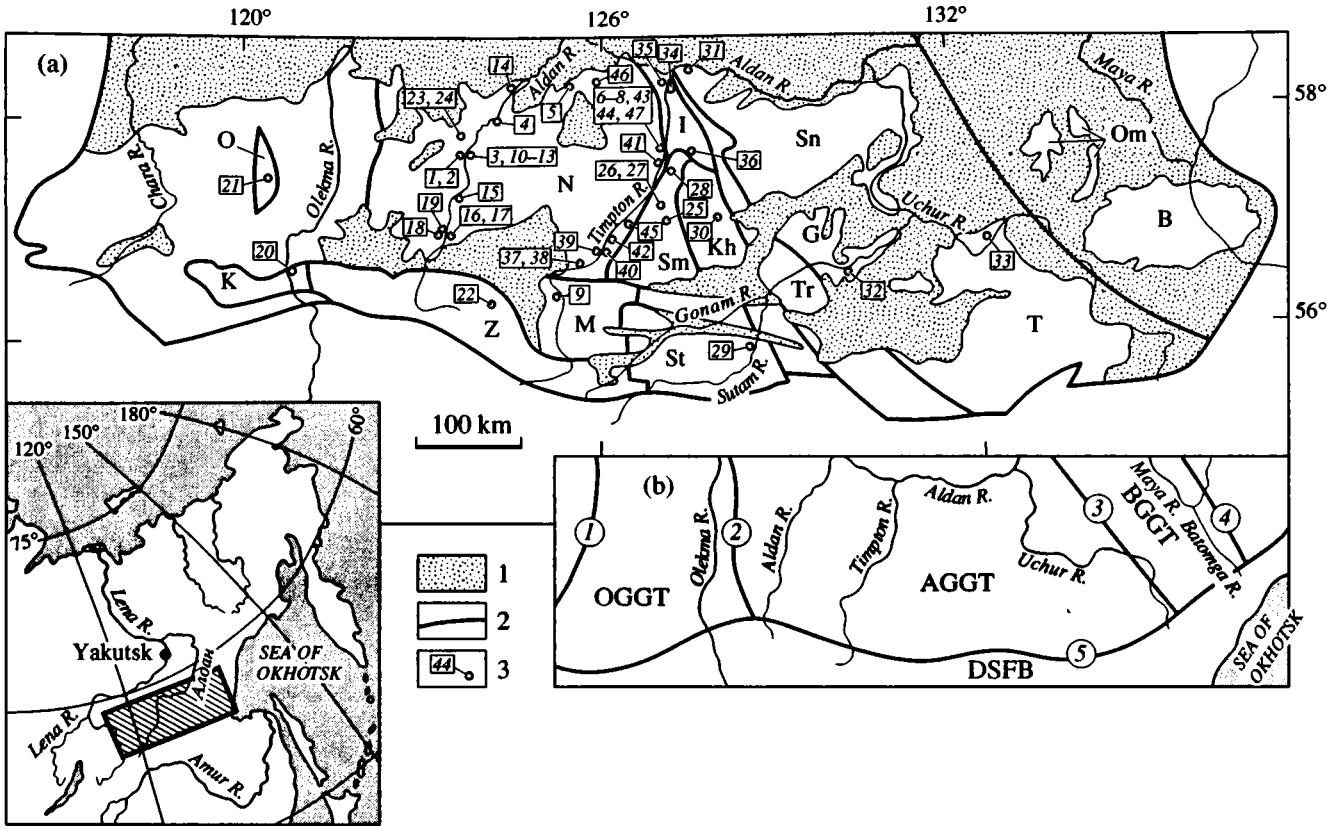

Fig. 1. Block structure (a) and tectonic zoning (b) of the Aldan Shield (after Popov and Smelov, 1996, and *Dokembriiskaya...*, 1988, respectively): (1) platform cover deposits; (2) block boundaries; (3) sampling sites numbered as in Tables 2–4. Tectonic blocks: (K) Kurulta; (O) Olomokit; (N) Nimnyr; (Z) Zverev; (M) Melemken; (I) Idzhek allochthon; (Sm) Seim; (St) Sutam; (Kh) Kholbolokh; (Tr) Tyrkanda; (Sn) Sunnagin; (G) Gonam; (T) Tyrkan; (Om) Omnya; (B) Batomga. Tectonic elements: (OGGT) Olekma granite–greenstone terrain; (AGGT) Aldan granulite–gneiss terrain; (BGGT) Batomga granite–greenstone terrain; (DSFB) Dzhugdzhur–Stanovoi fold belt. Fracture zones (encircled numbers): *1*, Zhuya; *2*, Amga; *3*, Ulkan; *4*, Nelkan; *5*, Stanovoi.

Precambrian Geology..., 1993). In accordance with this scheme, there were at least three formation stages of high-grade metamorphic supracrustal complexes of AGGT and adjacent Olekma granite–greenstone terrain (OGGT), all characterized in Table 1. Here, the so-called nonstratified infracrustal complex of tonalite-trondhjemite, granodiorite, and granite gneisses (i.e., plagio-, enderbite-, charnockite-, and granite-gneiss series) are considered as the basement of the supracrustal complexes.

The accumulation of predominantly terrigenous deposits of the Kurumkan and Amedichi groups and formation of volcanogenic–terrigenous rocks of the Zverev and Kurulta groups represent the first stage of development of the high-grade metamorphic supracrustal units. The Kurumkan Group sections are variable from one to another AGGT block (Table 1). In the western part of the Nimnyr block (Fig. 1), this group is composed of high-Al gneisses and quartzites in contrast to the eastern sections of the block, where gneisses of this type are main rock types. In the Melemken block (Fig. 1), identical gneisses and quartzites enclose lenses of calcareous-silicate rocks, calciphyres, and ferruginous and diopside quartzites (Table 1). Moreover, in the upper courses of the Aldan River near the village of Kurumkan, i.e., in the stratotype locality of the Kurumkan Group, quartzites, high-Al gneisses, basic crystalline schists, and amphibolites are closely associated with biotite to garnet–hypersthene-biotite gneisses and fine-banded diopside or scapolite-diopside plagiogneisses (Kitsul, 1989). Rocks of the last types were formerly considered as members of the Kurumkan Group (more precisely, of the Upper Aldan Formation of the Iengr Group, according to Dzevanovskii, 1958; Salop and Travin, 1974; Neelov and Mil'kevich, 1979; Frumkin, 1981); however, Dook and Pavlov (see *Rannii dokembrii ...*, 1986) excluded them, with due regard for data on chemical composition, from this unit. In the southwestern part of the Nimnyr block, these researchers (*Rannii dokembrii ...*, 1986) distinguished the allochthonous Amedichi Group as the age and facial analogue of the Kurumkan Group of the Melemken block (Table 1).

The Zverev Group of the synonymous AGGT block (Fig. 1) predominantly consists of basic crystalline schists associated with two-pyroxene and pyroxene–

Table 1. Main rock associations, their original composition, and model Sm–Nd ages of the high-grade metamorphic supracrustal complexes of the Aldan Shield (according to *Rannii dokembrii ...*, 1986, with additions and modifications)

Group	Tectonic block	Main rock association	Inferred original composition	$T_{Nd}(DM)$, Ga
Kurumkan	Nimnyr, western par	Quartzites, high-Al gneisses, Grt ± Opx–Bt plagioschists and gneisses, basic crystalline schists, Di, Di(± Scap) plagiogneisses	Ultrasilicites, aleuropelites and pelites; graywackes, tuffites or andesitic dacites; tholeiitic basalts, calcareous aleuropelites, carbonate–silicate rocks	3.5 2.9–3.0 2.3–2.4
	Nimnyr, eastern part	High-Al gneisses, basic crystalline schists	Pelites, hydromica subsialites, tholeiitic basalts, carbonates and carbonate–silicate rocks	2.9
	Melemken	Quartzites, high-Al gneisses, calciphyres, carbonate–silicate rocks, Fe-quartzites, basic crystalline schists	Ultrasilicites, hydromica subsialites, tholeiitic basalts, carbonates and carbonate–silicate rocks	3.1
Amedichi	Nimnyr, western part	Quartzites, high-Al gneisses, calciphyres, carbonate–silicate rocks, Fe-quartzites, basic crystalline schists	Continuous rock series from ultrasilicites to hydromica subsialites, tholeiitic basalts	3.0–3.3
Kurulta	Kurulta	Grt–Bt, Grt–Opx–Bt (± Sil, Crd) plagiogneisses, basic crystalline schists	siltstones of the graywacke type, tholeiitic basalts	3.0
	Olomokit			3.5
Zverev	Zverev	Basic crystalline schists, Opx–Cpx, Opx–Cpx–Hbl plagiogneisses, Grt–Bt plagiogneisse	Subalkaline gabbro–quartz diorites; siltstones of the graywacke type, tuffites, tholeiitic basalts	2.9
Chuga	Nimnyr, western part	Grt–Bt (± Sil), Bt, Bt–Hbl (± Di) plagiogneisses and basic crystalline schists, carbonate–silicate rocks, calciphyres, quartzites, high-Al gneisses	siltstones of the graywacke type, tuffites, silicites and ultrasilicites, tholeiitic and subalkaline basalts, subordinate andesites, carbonate and carbonate–silicate rocks	2.3–2.5
Seim	Seim	Grt–Bt plagiogneisses, gneisses, high-Al gneisses, Opx–Bt (± Grt), Opx–Cpx and Di–Hbl plagiogneisses, quartzites, carbonate–silicate rocks, basic and ultrabasic crystalline schists	Sandstones and siltstones of the graywacke type, tuffites, subsialites; picrites, picritic and tholeiitic basalts, their tuffs and tuffites	2.6–3.0
Kholbolokh	Kholbolokh	Grt–Bt (± Sil, Crd), Grt–Opx, Opx, Di–Opx plagiogneisses and gneisses, marbles, calciphyres, carbonate–silicate rocks, basic crystalline schists	Sandstones and siltstones of the graywacke type, silty shales, tuffites, tholeiitic basalts, carbonate and carbonate–silicate rocks	2.2
	Sunnagin			2.6
	Gonam			2.3
	Tyrkan			2.1
Kyurikan	Kholbolokh	Bt, Grt–Bt, Opx–Bt (±Grt), Di (± Opx, Hbl) plagiogneisses, Opx (± Di, Hbl), Di (± Hbl) plagioschists, carbonate–silicate rocks, calciphyres, marbles, Opx–Cpx–Hbl gneisses and basic crystalline schists	Siltstones of the graywacke type and/or tuffites, aleuropelites, carbonate and carbonate–silicate rocks, subalkaline basalts, basaltic andesites, and andesites	2.3–2.4
Idzhek	Idzhek allochthon	Opx (± Di, Hbl), Grt–Opx–Bt plagiogneisses, gneisses, Di (±Scap), Di–Hbl plagioschists, carbonate–silicate rocks, quartzites, basic crystalline schists, Opx–Cpx–Bt and Opx–Bt gneisses	Carbonate siltstones, carbonate to silty shales, siltstones of the graywacke type, subalkaline basalts and andesites	2.4
Fedorov	Nimnyr	Hbl, Bt–Hbl, Di–Hbl, Opx–Cpx–Hbl, and Opx–Cpx plagiogneisses, basic and ultrabasic crystalline schists, Grt–Bt gneisses, Di and Phl–Di rocks, marbles, calciphyres	Subalkaline, calc-alkaline and alkaline basalts, andesites, dacites and rhyodacites, tuffs and tuffites, carbonate and carbonate–silicate rocks	2.2–2.4 3.2

amphibole plagiogneisses corresponding in composition to subalkaline basalts and andesites and/or to gabbro and diorites of the same type. Garnet–biotite plagiogneisses are scarce here. The characteristic rock association of the Kurulta Group (Kurulta and Olomokit blocks of OGGT, Fig. 1) is represented by garnet–biotite and garnet hypersthene–biotite (± sillimanite and cordierite) plagiogneisses coexisting with basic crystalline schists.

The second evolutionary stage of high-grade metamorphic supracrustal complexes of the central Aldan

Shield resulted in accumulation of volcanic, carbonate, and terrigenous deposits of the Chuga (western part of the Nimnyr block), Seim (Seim and Sutam blocks), and Kholbolokh groups (Kholbolokh, Sunnagin, Gonam, Tyrkan, and Tyrkanda blocks, see Fig. 1). Typical rocks of the Chuga Group are garnet–biotite (± sillimanite), biotite, and biotite–amphibole (± diopside) plagiogneisses, basic crystalline schists, carbonate–silicate rocks, calciphyres, quartzites, and high-Al gneisses. The Seim Group is composed of plagiogneisses and gneisses of the garnet–biotite, high-Al, hypersthene–biotite (± garnet), two-pyroxene, and diopside–amphibole types associated with crystalline schists of basic and ultrabasic compositions. Supracrustal rocks of the Kholbolokh Group are represented by garnet–biotite (± sillimanite and cordierite); garnet–hypersthene; hypersthene; diopside–hypersthene plagiogneisses and gneisses; and also by marbles, calciphyres, carbonate–silicate rocks, and basic crystalline schists.

The Kyurikan (Kholbolokh, Gonam, and Tyrkan blocks) and Idzhek (Nimnyr block) groups of volcanic, carbonate, and terrigenous rocks, together with the Fedorov Group (Nimnyr block) predominantly volcanic in composition, exemplify units of the third evolutionary stage of the supracrustal complexes (Table 1). The Kyurikan Group consists of biotite, garnet–biotite, hypersthene–biotite (± garnet), and diopside plagiogneisses associated with hypersthene (± diopside and amphibole) plagioschists, carbonate–silicate rocks, calciphyres, marbles, two-pyroxene–amphibole gneisses, and basic crystalline schists. The Idzhek Group is recognized in a tectonic nappe (Idzhek allochthon) thrusted over the northeastern and northern parts of the Nimnyr and Seim blocks, respectively (*Rannii dokembrii* ..., 1986). This unit includes hypersthene (± diopside and amphibole) and garnet–hypersthene–biotite plagiogneisses or gneisses, diopside (± scapolite) and diopside–amphibole plagioschists, carbonate–silicate rocks, quartzites, basic crystalline schists, and two-pyroxene–biotite or hypersthene–biotite gneisses. Rocks of the Fedorov Group are exposed in central and eastern areas of the Nimnyr block, where they occur, like rocks of the Idzhek Group, in the allochthonous position. Dook (1994) considered them as members of the "Fedorov nappe" originating before the Idzhek allochthon and Timpton thrust formation. Principal rocks of the Fedorov Group are amphibole, biotite–amphibole, diopside–amphibole, two-pyroxene–amphibole, or two-pyroxene plagiogneisses and also basic to ultrabasic crystalline schists, diopside or phlogopite–diopside rocks, marbles, and calciphyres.

After the accumulation of the Kyurikan, Idzhek, and Fedorov group, supracrustal rocks of AGGT and eastern OGGT experienced the recurrent structural transformation and metamorphism of the granulite or transitional granulite–amphibolite facies (*Rannii dokembrii* ..., 1986; *Dokembriiskaya* ..., 1988; *Precambrian Geology* ..., 1993). The latest metamorphic events took place between 1962 ± 5 and 1907 ± 15 Ma in the northwestern part of the Nimnyr block (Kotov *et al.*, 1993) and between 1993 ± 1 and 1916 ± 10 Ma in its eastern area (Bibikova *et al.*, 1986; Kotov *et al.*, 1995).

In order to complete the brief characteristics of supracrustal rocks in the central area of the Aldan Shield, we would like to stress that their time constraints are evaluated thus far on the basis of indirect evidence, such as geological relations with structural elements and magmatic intrusions different in age. Special structural and geochronological investigations of recent time revealed the following:

(1) Rocks of the Fedorov Group experienced the earliest structural transformation and metamorphism of the granulite facies 1993 ± 1 Ma ago (Kotov *et al.*, 1995).

(2) In northwestern part of the Nimnyr block, supracrustal rocks of the Kurumkan and Chuga groups display structural patterns of plastic thrusts formed not before ~2.4 Ga (Sal'nikova *et al.*, 1996).

(3) Supracrustal rocks of the Seim Group were affected by structural deformations and granulite metamorphism related to formation of the deep-seated Timpton thrust, which appears to be no older than 1950 ± 19 Ma (Kotov *et al.*, 1995).

As one can see, the geochronological data available at present elucidate to some extent only the minimum age limit of certain supracrustal complexes of the Aldan Shield. In any case, this limit is marked by manifestation of the Early Proterozoic endogenic processes.

INVESTIGATION METHODS

One of possible ways to apprise age constraints of high-grade supracrustal complexes is to study the Nd isotope composition in their metasedimentary clastic rocks. Because of the relatively constant $^{147}Sm/^{144}Nd$ ratios in siliciclastic rocks (~0.11 ± 0.02), which differ considerably from those in the mantle (≥0.2), and due to a negligible effect of weathering, transportation, diagenesis, and metamorphism on the values of those ratios (DePaolo, 1988; Taylor and McLennan, 1985), the study on Nd isotope composition in such rocks offers a possibility to infer the *mean*[1] model age of their provenances (McCulloch and Wasserburg, 1978; O'Nions *et al.*, 1983; Miller *et al.*, 1986) and to define thus the maximum age limit of supracrustal complexes. Another independent way to obtain information of this kind is the study on Sm–Nd isotope systematics for calc-alkaline and alkaline metavolcanic rocks[2] associated with metasediments, when we are able to deter-

[1] It should be stressed that the model Nd age of sediments, whose material is derived from several source areas, indicates the weighted-mean crustal residence time of Nd derived from the latter.

[2] Volcanic rocks of the tholeiite series are inappropriate for the correct determination of model ages, because their $^{147}Sm/^{144}Nd$ ratios are high (≥0.15) and close to those characterizing the chondrite undifferentiated reservoir (CHUR) or the depleted mantle (DM).

mine, to a first approximation, the segregation time of parental melts from the mantle (DePaolo, 1988).

In order to take into account probable fractionation of Sm and Nd in the course the crustal processes, we also calculated the two-stage model ages T_{Nd}(DM-2st) following method by Liew and Hofmann (1988). In this case, the last event of granulite metamorphism in the AGGT (~1960 Ma ago) was taken for the minimum age limit of the fractionation.

Rock samples for Sm–Nd isotopic study were selected on the basis of geological and geochemical criteria used to reveal an affinity between them and protoliths of clastic sediments or calc-alkaline volcanic rocks. Parental rock series for the high-grade metamorphic supracrustal complexes were reconstructed by using classification diagrams by Neelov (1980), Predovskii (1980), Dennen and Moore (1971), and discriminant function by Shaw (1972). In this case, we also took into account the earlier results of lithological and geochemical studies (Kulish, 1973; Travin, 1977; Neelov and Mil'kevich, 1979; Reutov, 1981; *Rannii dokembrii ...*, 1986; Velikoslavinskii *et al.*, 1990).

Sampling sites are shown in Fig. 1. Data on mineral and chemical composition of most samples are presented in Tables 2 and 3, and their affinity with various types of sediments is indicated in Fig. 2.

The original sedimentary genesis is confidently defined for the high-Al gneisses and schists usually bearing sillimanite and cordierite (± garnet), and corresponding in chemical composition to clayey rocks, or to pelites in a broad sense. The garnet–biotite plagiogneisses and gneisses, including their hypersthene-bearing varieties, are quite similar in their chemical composition to graywackes, although they, like hypersthene–biotite and biotite plagiogneisses, may also be compatible with tuffites or intermediate and acid calc-alkaline magmatic rocks. Basic crystalline schists and hypersthene-bearing rocks of the Fedorov Group apparently represent the basic to acid metavolcanics (*Rannii dokembrii ...*, 1986; Velikoslavinskii *et al.*, 1990).

Analysis of Sm and Nd isotope composition was performed at the Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences, using the procedure described before (Neymark *et al.*, 1993; Kotov *et al.*, 1995). The total blank during measurements was 0.03–0.2 ng for Sm and 0.1–0.5 ng for Nd. Measured $^{143}Nd/^{144}Nd$ ratios were normalized to $^{146}Nd/^{144}Nd = 0.7219$. The analytical precision is ±0.5% (2σ) for Sm and Nd concentrations, and ±0.005% for the $^{143}Nd/^{144}Nd$ ratio. For 11 measurements in the La Jolla standard, the weighted-mean $^{143}Nd/^{144}Nd$ ratio is 0.511894 ± 8 (2σ).

Calculating the $\varepsilon_{Nd}(0)$ values and model T_{Nd}(DM) ages, we used isotopic parameters of the CHUR ($^{143}Nd/^{144}Nd = 0.512638$ and $^{147}Sm/^{144}Nd = 0.1967$; Jacobsen and Wasserburg, 1984) and DM ($^{143}Nd/^{144}Nd = 0.513151$ and $^{147}Sm/^{144}Nd = 0.2136$; Goldstein and Jacobsen, 1988). The two-stage model ages T_{Nd}(DM-2st) are calculated on the basis of the mean isotopic ratio in the crust $^{147}Sm/^{144}Nd = 0.12$ (Taylor and McLennan, 1985).

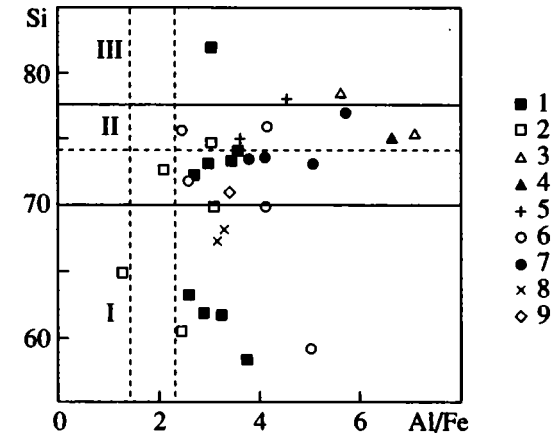

Fig. 2. Si–Al/Fe diagram (Dennen and Moore, 1971) for studied samples from the Kurumkan (1), Amedichi (2), Kurulta (3), Zverev (4), Chuga (5), Seim (6), Kholbolokh (7), Kyurikan (8), and Idzhek (9) groups of supracrustal rocks: (I) pelites; (II) graywackes; (III) subgraywackes (dashed lines restrict the overlapping intervals of pelites and graywackes).

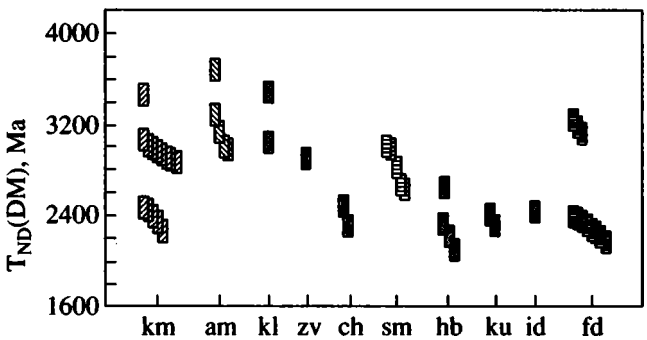

Fig. 3. Values of model T_{Nd}(DM) ages for supracrustal rocks from the Kurumkan (km), Amedichi (am), Kurulta (kl), Zverev (zv), Chuga (ch), Seim (sm), Kholbolokh (hb), Kyurikan (ku), Idzhek (id), and Fedorov (fd) groups of the central Aldan Shield.

RESULTS OF Sm–Nd ISOTOPIC STUDIES

The Sm–Nd isotopic parameters of supracrustal rocks of the *Kurumkan Group* are studied in samples of high-Al gneisses (metapelites) from different sections, and also in garnet ± hypersthene–biotite plagiogneisses (metagraywackes) sampled in its stratotype. The Early Archean model age[3] T_{Nd}(DM) = 3.5 Ga is inferred for one sample of the high-Al gneiss (Table 4, Fig. 3) spatially interrelated with the oldest tonalite–trondhjemite gneisses of the nonstratified infracrustal complex of the

[3] In this paper, we use the subdivision scheme of the Early Precambrian authorized in *Resheniya ...*, (1992).

Table 2. Stratigraphic level and mineral composition of high-grade metamorphic supracrustal rocks of the Aldan Shield

No.	Sample no.	Group	Mineral composition
1	B-1906/10	Kurumkan	Sillimanite–cordierite gneiss
2	B-1906/21	"	Garnet–sillimanite–cordierite–biotite gneiss
3	Dr-0316/12	"	Cordierite–garnet–biotite gneiss
4	Dr-0261v	"	Garnet–cordierite gneiss
5	B-2893	"	Cordierite gneiss
6	N-1/1	"	Cordierite–garnet–hypersthene gneiss
7	N-4/2	"	Sillimanite–cordierite–hornblende gneiss
8	K-188	"	Garnet–biotite plagiogneiss
9	V-31b	"	Garnet–sillimanite gneiss
10	B-1907/6	"	Garnet–biotite plagiogneiss
11	B-1907/6a	"	Garnet–hypersthene–biotite plagiogneiss
12	B-1907/17	"	Biotite–hypersthene plagiogneiss
13	B-1907/18a	"	Biotite–hypersthene plagiogneiss
14	B-1922/1	"	Garnet–hypersthene–biotite plagiogneiss
15	B-1108/1	Amedichi	Garnet–biotite–sillimanite–cordierite gneiss
16	B-1815/11	"	Garnet–biotite plagiogneiss
17	B-1817	"	Garnet–biotite plagiogneiss
18	B-3298/2	"	Garnet–biotite plagiogneiss
19	B-762-1a	"	Garnet–biotite–cordierite–sillimanite gneiss
20	R-5029/6	Kurulta	Garnet–hypersthene–biotite plagiogneiss
21	B-2635/3	"	Garnet–biotite plagiogneiss
22	Z-33/3	Zverev	Garnet–biotite gneiss
23	B-3260/1	Chuga	Biotite plagiogneiss
24	B-3274	"	Garnet–biotite plagiogneiss
25	Z-507/4	Seim	Garnet–hypersthene–biotite plagiogneiss
26	Z-511/1	"	Hypersthene–biotite plagiogneiss
27	Z-511/11	"	Garnet–hypersthene plagiogneiss
28	LB-200/6	"	Garnet–biotite plagiogneiss
29	B-614-9a	"	Garnet–biotite–sillimanite–hypersthene gneiss
30	D-1116/4	Kholbolokh	Biotite–hypersthene gneiss
31	B-2845/11	"	Garnet–biotite plagiogneiss
32	Z-305/1	"	Biotite–hypersthene gneiss
33	PK-310/9	"	Biotite plagiogneiss
34	B-1963	Kyurikan	Garnet–cordierite–biotite gneiss
35	Ki-13	"	Garnet–biotite–sillimanite gneiss
36	D-82/2	Idzhek	Garnet–biotite plagiogneiss
37	1822	Fedorov	Crystalline amphibole–clinopyroxene-plagioclase schist
38	1826	"	Crystalline amphibole–clinopyroxene-plagioclase schist
39	1862	"	Crystalline amphibole–diopside–hypersthene-plagioclase schist
40	1860a	"	Crystalline amphibole–diopside–hypersthene-plagioclase schist
41	1890	"	Biotite–amphibole–hypersthene gneiss
42	1866	"	Biotite–hypersthene plagiogneiss
43	1908b	"	Biotite–hypersthene plagiogneiss
44	1907a	"	Biotite–amphibole– hypersthene gneiss
45	1880g	"	Biotite–amphibole–diopside–hypersthene gneiss
46	B-2880/1	"	Hypersthene–amphibole plagiogneiss
47	B-132	"	Crystalline amphibole–clinopyroxene–plagioclase schist

Note: Letter symbols mark samples from collections of V.I. Berezkin (B), V.S. Shkodzinskii (V), V.L. Dook (D), G.M. Drugova (Dr), A.M. Zedgenizov (Z), A.B. Kotov (K), V.I. Kitsul (Ki), N.Yu. Zagornaya (N), P.A. Kopylov (PK), and S.S. Rozhin (R); samples V-132 and others without indices are from collection of S.D. Velikoslavinskii.

Table 3. Representative chemical analyses (wt %) of high-grade metamorphic supracrustal rocks of the Aldan Shield

Group	Kurumka									Amedichi				
Sample no.	B-1906/10	B-1906/21	B-2893	V-31b	B-1907/6	B-1907/6a	B-1907/17	B1907/18a	B1922/1	B-1108/1	V762-1a	B-1815/11	B-1817	B-3298/2
Ordinal no.	1	2	5	9	10	11	12	13	14	15	16	17	18	19
SiO_2	53.40	57.27	54.31	58.47	75.06	63.27	62.58	62.02	66.70	56.21	62.62	58.04	64.06	66.46
TiO_2	1.05	1.10	1.01	1.14	0.57	0.88	0.71	0.73	0.78	0.82	1.30	0.72	1.12	0.77
Al_2O_3	25.40	22.03	21.67	20.61	10.49	15.06	14.50	14.62	15.39	21.91	17.21	14.53	13.69	14.32
Fe_2O_3	5.17	0.96	3.13	1.05	0.60	0.83	0.65	0.47	0.01	4.46	0.84	4.19	3.39	7.46
FeO	5.03	10.06	6.75	10.50	4.38	5.52	6.35	7.34	6.16	8.85	7.20	13.04	6.38	–
MnO	0.08	0.10	0.10	0.07	0.06	0.07	0.10	0.10	0.03	0.08	0.09	0.66	0.11	0.09
MgO	1.66	2.45	2.70	2.60	1.80	4.22	4.21	4.33	2.02	2.29	2.04	5.07	2.44	1.92
CaO	0.45	1.33	1.64	1.34	1.51	3.88	4.20	3.95	1.84	0.27	1.80	0.54	2.03	2.63
Na_2O	1.68	1.43	3.26	1.03	2.40	2.73	3.01	2.90	5.03	0.27	2.30	0.01	3.44	3.35
K_2O	5.31	2.16	4.06	2.54	1.85	1.79	2.19	2.09	1.76	1.99	2.28	1.86	2.19	2.39
P_2O_5	0.27	0.14	0.11	0.08	0.10	0.21	0.21	0.18	0.06	0.26	0.08	0.04	0.16	0.24
L.O.I.	1.06	1.17	1.39	0.59	0.40	0.99	1.06	1.00	0.28	3.21	1.92	2.03	1.17	0.14
Total	100.56	100.20	100.13	100.02	99.22	99.45	99.77	99.73	100.06	100.62	99.68	100.73	100.18	99.77
Group	Kurulta		Zverev	Chuga		Seim					Kholbolokh			
Sample no.	R-5029/6	B-2635/3	Z-33/3	B-3260/1	B-3274	Z-507/4	Z-511/1	Z-511/11	LB-200/6	V-614-9‡	D-1116/4	B-2845/11	Z-305/1	PK-310/9
Ordinal no.	20	21	22	23	24	25	26	27	28	29	30	31	32	33
SiO_2	70.85	68.31	67.69	69.80	66.16	60.98	67.19	68.33	63.93	53.33	64.62	66.65	68.85	64.37
TiO_2	0.37	0.54	0.53	0.43	0.72	0.84	0.40	0.90	0.77	1.03	0.71	0.70	0.33	0.57
Al_2O_3	13.92	16.52	16.52	13.82	14.70	17.81	14.50	13.15	15.15	25.91	15.55	16.26	14.81	16.71
Fe_2O_3	0.44	0.65	0.92	1.70	6.35	1.29	1.38	–	2.26	0.53	0.44	1.68	4.08	5.20
FeO	3.12	2.72	2.70	2.75	–	5.00	3.72	7.69	6.42	6.87	5.48	4.12	–	–
MnO	0.04	0.02	0.04	0.04	0.06	0.06	0.11	0.11	0.04	0.01	0.06	0.11	0.06	0.05
MgO	2.53	1.16	1.53	1.53	2.86	2.26	1.88	2.67	3.80	5392	2.60	1.55	1.16	2.19
CaO	3.85	2.91	2.30	0.64	3.30	4.70	3.65	3.32	1.64	0.44	3.37	2.58	3.01	3.18
Na_2O	3.00	4.71	2.87	2.42	2.23	3.90	2.81	1.69	3.19	1.60	2.35	2.77	3.20	4.50
K_2O	1.15	2.04	4.06	3.38	2.49	1.51	3.04	0.45	2.53	3.15	0.21	3.61	3.32	1.75
P_2O_5	0.01	0.13	0.06	0.15	0.16	0.08	0.07	0.03	0.01	0.06	0.01	0.14	0.07	0.16
L.O.I.	0.41	0.74	0.72	0.84	0.84	0.48	1.23	1.71	0.58	1.09	–	0.20	0.60	0.81
Total	99.69	100.45	99.94	97.50	99.87	98.91	99.98	100.05	100.32	99.94	95.40	100.37	99.49	99.49
Group	Kyurikan		Idzhek	Fedorov										
Sample no.	B-1963	K-13	D-82/2	1822	1826	1862	1860a	1890	1866	1908b	1907a	1880g	B-2880/1	V-132
Ordinal no.	34	35	36	37	38	39	40	41	42	43	44	45	46	47
SiO_2	60.33	62.40	62.28	47.08	49.37	49.08	54.50	60.76	65.11	66.38	72.98	75.10	68.02	45.34

Table 3. (Contd.)

Group	Kurumkan		Idzhek	Fedorov										
Sample no.	B-1963	K-13	D-82/2	1822	1826	1862	1860a	1890	1866	1908b	1907a	1880g	B-2880/1	V-132
Ordinal no.	34	35	36	37	38	39	40	41	42	43	44	45	46	47
TiO_2	0.95	0.81	0.77	1.76	1.06	1.10	1.00	0.48	0.49	0.39	0.27	0.21	0.46	1.00
Al_2O_3	18.94	19.03	16.48	14.18	13.16	16.24	16.35	16.24	16.83	17.67	15.05	13.34	14.79	13.38
Fe_2O_3	0.01	1.85	2.86	9.23	7.27	6.34	5.64	3.56	4.15	2.06	1.60	2.00	1.02	4.01
FeO	8.48	6.51	4.38	7.38	7.17	5.30	3.99	4.67	2.57	2.42	1.42	0.80	4.02	7.62
MnO	0.04	0.06	0.12	0.26	0.31	0.21	0.12	0.15	0.15	0.06	0.07	0.05	0.09	0.19
MgO	3.39	3.16	3.38	5.16	6.77	6.02	4.33	2.97	1.77	1.70	0.71	0.91	1.40	10.76
CaO	2.28	0.95	4.07	9.68	8.94	8.41	7.29	5.98	4.46	4.24	2.25	3.04	4.17	11.68
Na_2O	2.95	1.92	4.06	3.42	3.34	3.98	3.20	3.20	4.06	4.06	3.54	3.22	3.43	2.34
K_2O	2.19	2.74	1.34	0.89	1.09	1.43	1.89	1.00	0.82	1.05	2.53	1.55	2.02	2.29
P_2O_5	0.06	0.10	0.01	0.10	0.05	0.34	0.47	0.13	0.17	0.15	0.05	–	–	0.48
L.O.I.	1.29	0.72	0.24	0.81	1.17	0.98	0.68	0.71	0.70	0.43	0.38	0.69	0.65	1.32
Total	100.88	100.25	99.99	99.95	99.70	99.43	99.46	99.85	101.28	100.61	100.85	100.91	100.07	100.41

Note: Ordinal numbers correspond to those in Tables 2 and 4, and in Fig. 1.

northwestern AGGT. The latter are ca. 3.3 Ga old (Nutman *et al.*, 1992) and show the model Nd-age interval of 3.5–3.9 Ga (Sal'nikova *et al.*, 1996). It is likely that precisely these tonalite–trondhjemite gneisses served as the source rocks for the high-Al gneiss protoliths in the region.

All other samples from the Kurumkan Group yielded, disregarding the type and geographic position of a section, the model Nd ages ranging in narrow intervals: $T_{Nd}(DM)$ = 2.9–3.1 and $T_{Nd}(DM\text{-}2st)$ = 3.0–3.2 Ga (Table 4, Fig. 3). This may indicate that original clastic material of high-Al gneisses of the Kurumkan Group was derived from provenances composed of the Upper Archean rocks and accumulated no earlier than ~2.9 Ga ago.

$T_{Nd}(DM)$ and $T_{Nd}(DM\text{-}2st)$ values for the garnet ± hypersthene–biotite plagiogneisses of the Kurumkan Group are 2.3–2.4 and 2.3–2.5 Ga, respectively (Table 4, Fig. 3). If their protoliths were graywackes, then these values of model Nd ages illustrate the Early Proterozoic age of the corresponding source areas. In other case, when the protoliths were either tuffites or magmatic rocks, the calculated model ages mark the crust-forming event[4] that occurred ≤2.3 Ga ago. Thus, disregarding the original nature of the garnet ± hypersthene–biotite plagiogneisses of the Kurumkan Group, their protoliths were formed at the Early Proterozoic stage of the AGGT evolution.

The model Nd ages for the high-Al gneisses (metapelites) and garnet-biotite plagiogneisses (metagraywackes) of the *Amedichi Group* are comparable: $T_{Nd}(DM)$ = 3.0–3.1 and 3.0–3.3 Ga, and $T_{Nd}(DM\text{-}2st)$ = 3.2 and 3.1–3.2 Ga, respectively, for the indicated rock types (Table 4, Fig. 3). The only exception is Sample B-1817 of the garnet–biotite plagiogneiss that yielded the model age $T_{Nd}(DM)$ = 3.7 Ga. This sample also revealed, in contrast to the high-Al gneiss from the Kurumkan Group yielding an Early Archean model age, the higher value of $^{147}Sm/^{144}Nd \approx 0.15$. Its mineral and chemical compositions (Tables 2 and 3) offer some ground to consider this value of the Sm/Nd ratio as reflecting the sample enrichment in garnet in the course of metamorphic differentiation. The two-stage model age $T_{Nd}(DM\text{-}2st)$ for this garnet–biotite plagiogness is 3.1 Ga, i.e., concordant with model ages of other rock samples from the Amedichi Group. Accordingly, sediments of the group accumulated during the erosion of rocks, whose model age is Late Archean in average.

The hypersthene–garnet–biotite plagiogneiss (metagraywacke) of the *Kurulta Group* sampled in the Kurulta OGGT block is characterized by the values $T_{Nd}(DM)$ = 3.0 and $T_{Nd}(DM\text{-}2st)$ = 3.3 Ga (Table 4), whereas the garnet–biotite plagiogneiss of the same group from the Olomokit OGGT block yielded the val-

[4] We would like to stress that the term "crust-forming event" means the time interval when the crustal rocks were initially isolated from the mantle source. Its age, in contrast to the age of crystallization or thermal and orogenic transformation of the crustal rocks, can be calculated as model and corresponding to the time when the Nd isotopic composition in crustal rocks was identical to that in the mantle source (Kovalenko *et al.*, 1996, p. 702).

Table 4. Sm–Nd isotopic data for high-grade metamorphic supracrustal rocks of the Aldan Shield

Ordinal no.	Sample no.	Sm, μg/g	Nd, μg/g	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ ($\pm 2\sigma_{meas}$)	$\varepsilon_{Nd}(0)$	$T_{Nd}(DM)$, Ma	$T_{Nd}(DM\text{-}2st)$, Ma
1	B-1906/10	8.04	46.18	0.1056	0.511030 ± 8	−31.4	2975	3130
2	B-1906/21	8.31	49.06	0.1027	0.510944 ± 9	−33.0	3013	3207
3	Dr-0316/12	3.25	17.04	0.1156	0.510912 ± 9	−33.7	3456	3525
4	Dr-0261v	8.32	47.68	0.1058	0.511029 ± 16	−31.4	2980	3135
5	B-2893	8.01	50.62	0.0960	0.510912 ± 9	−33.7	2885	3121
6	N-1/1	9.33	57.53	0.0984	0.500978 ± 8	−32.4	2856	3063
7	N-4/2	6.44	36.51	0.1070	0.511106 ± 8	−29.9	2904	3036
8	K-188	6.37	37.60	0.1027	0.511009 ± 10	−31.8	2925	3103
9	V-31·	4.53	22.66	0.1213	0.511286 ± 10	−26.4	3060	3044
10	B-1907/6	7.33	41.53	0.1071	0.511429 ± 16	−23.6	2453	2520
11	B-1907/6a	4.74	24.72	0.1162	0.511588 ± 10	−20.5	2435	2454
12	B-1907/17	3.49	19.22	0.1097	0.511519 ± 6	−21.8	2383	2429
13	B-1907/18a	3.65	19.83	0.1113	0.511580 ± 4	−20.6	2331	2365
14	B-1922/1	6.79	32.58	0.1105	0.511618 ± 9	−19.9	2258	2288
15	B-1108/1	8.69	50.98	0.1034	0.510979 ± 9	−32.4	2985	3166
16	B-1815/11	3.91	21.77	0.1090	0.511072 ± 8	−30.5	3008	3131
17	B-1817	3.58	14.51	0.1496	0.511588 ± 6	−20.5	3687	3144
18	B-3298/2	6.38	31.27	0.1236	0.511193 ± 4	−28.2	3292	3241
19	V-762-1a	8.05	43.17	0.1131	0.511067 ± 6	−30.6	3139	3225
20	R-5029/6	1.10	6.96	0.0960	0.510786 ± 7	−36.1	3043	3320
21	B-2635/3	3.19	17.23	0.1124	0.510817 ± 6	−35.5	3488	3611
22	Z-33/3	8.23	43.73	0.1141	0.511252 ± 10	−27.0	2950	3008
23	B-3260/1	5.97	32.15	0.1126	0.511621 ± 5	−19.8	2298	2325
24	B-3274	5.54	29.07	0.1156	0.511552 ± 4	−21.2	2474	2498
25	Z-507/4	8.6	66.12	0.0787	0.510789 ± 5	−36.1	2654	2959
26	Z-511/1	2.78	14.92	0.1130	0.511416 ± 5	−23.8	2616	2664
27	Z-511/11	1.65	8.39	0.1191	0.511395 ± 7	−24.2	2814	2823
28	LB-200/6	2.81	14.87	0.1145	0.511194 ± 5	−28.8	2990	3050
29	V-614-9a	9.08	53.52	0.1029	0.510956 ± 6	−32.8	3001	3191
30	D-1116/4	4.69	25.26	0.1123	0.511675 ± 4	−18.8	2212	2233
31	B-2845/11	5.93	30.75	0.1166	0.511464 ± 6	−22.9	2637	2662
32	Z-305/1	1.79	11.78	0.0921	0.511301 ± 7	−26.1	2311	2415
33	PK-310/9	3.75	26.09	0.0871	0.511409 ± 5	−24.0	2092	2138
34	B-1963	7.36	41.55	0.1074	0.511539 ± 9	−21.4	2303	2349
35	Ki-13	6.54	35.39	0.1120	0.511545 ± 12	−21.3	2398	2435
36	D-82/2	3.85	24.15	0.0967	0.511288 ± 9	−26.3	2418	2532
37	1822	4.03	15.4	0.1586	0.511996 ± 13	−12.5	3177	–
38	1826	3.62	15.21	0.1444	0.511669 ± 11	−18.9	3238	–
39	1862	8.86	57.62	0.0933	0.511269 ± 9	−26.7	2373	–
40	1860a	7.19	41.94	0.1040	0.511438 ± 5	−23.4	2371	–
41	1890	7.58	39.87	0.1153	0.511693 ± 7	−18.4	2252	2267
42	1866	6.22	34.00	0.1110	0.511544 ± 14	−21.3	2375	2415
43	1908b	1.42	9.18	0.0937	0.511365 ± 5	−24.8	2261	2345
44	1907a	2.37	17.06	0.0843	0.511313 ± 4	−25.8	2158	2234
45	1880g	1.92	12.07	0.0966	0.511367 ± 5	−24.8	2313	2402
46	B-2880/1	3.38	17.22	0.1192	0.511776 ± 5	−16.8	2211	2213
47	V-132	4.05	15.37	0.1599	0.512045 ± 5	−11.6	3116	–

Note: Ordinal numbers correspond to those in Tables 2 and 3, and in Fig. 1. Sample positions: (1–8, 10–14) Kurumkan Group, Nimnyr block; (9) Kurumkan Group, Melemken block; (15–19) Amedichi Group, Nimnyr block; (20) Kurulta Group, Kurulta block; (21) Kurulta Group, Olomokit block; (22) Zverev Group, Zverev block; (23, 24) Chuga Group, Nimnyr block; (25–28) Seim Group, Seim block; (29) Seim Group, Sutam block; (30) Kholbolokh Group, Kholbolokh block; (31) Kholbolokh Group, Sunnagin block; (32) Kholbolokh Group, Gonam block; (33) Kholbolokh Group, Tyrkanda block; (34, 35) Kyurikan Group, Kholbolokh block; (36) Idzhek Group, Idzhek allochthon; (37–47) Fedorov Group, Nimnyr block.

ues 3.5 and 3.6 Ga, respectively. This means that in the last case the provenance for protolith was predominantly composed of older Lower Archean rocks, although the Late Archean formation time cannot be excluded for the rocks of both blocks. Nevertheless, the Early Archean model age of ~3.7 Ga was also calculated for the tonalite–gneiss sample from the nonstratified infracrustal OGGT complex exposed near the eastern boundary of the Olomokit block (Neymark *et al.*, 1993).

$T_{Nd}(DM)$ and $T_{Nd}(DM\text{-}2st)$ values for the garnet–biotite gneiss (metagraywacke) of the *Zverev Group* sampled in tectonic block of the same name are 2.9 and 3.0 Ga, respectively (Table 4), which is evidence of the Late Archean model age of the related provenance.

One- and two-stage model ages for biotite and garnet–biotite plagiogneisses (metagraywackes) of the *Chuga Group* are within the interval of 2.3–2.5 Ga (Table 4, Fig. 3). These and geological data suggest this group to be formed during the Early Proterozoic evolutionary stage of the Aldan Shield.

Garnet ± hypersthene ± biotite plagiogneisses and gneisses (metagraywackes) of the *Seim Group*, which were sampled in the Seim block, show a wide range (2.6–3.0 Ga) of model $T_{Nd}(DM)$ ages (Fig. 3) that are very close to their $T_{Nd}(DM\text{-}2st)$ ages of 2.7–3.0 Ga (Table 4). Model ages $T_{Nd}(DM)$ = 3.0 and $T_{Nd}(DM\text{-}2st)$ = 3.2 Ga characterizing the garnet–biotite–sillimanite–hypersthene gneiss (metapelite) of the same group in the Sutam block are also close to this interval, and protoliths of this group apparently were derived from provenances attributable to the Late Archean terrain in general.

$T_{Nd}(DM)$ values for biotite–hypersthene and biotite plagiogneisses and gneisses (metagraywackes) of the *Kholbolokh Group* (Kholbolokh, Gonam, and Tyrkan blocks of the AGGT, Fig. 1) range from 2.1 to 2.3 and $T_{Nd}(DM\text{-}2st)$ = 2.1–2.4 Ga (Table 4), whereas the garnet–biotite plagiogneiss (metagraywacke) from the Sunnagin block yields: $T_{Nd}(DM)$ = 2.6 and $T_{Nd}(DM\text{-}2st)$ = 2.7 Ga (Fig. 3). Similar values of $T_{Nd}(DM)$ = 2.3–2.4 and $T_{Nd}(DM\text{-}2st)$ = 2.4 Ga are calculated for the gneisses of garnet–cordierite–biotite and the garnet–biotite–sillimanite composition (metapelites) from the *Kyurikan Group*, and the garnet–biotite plagiogneiss (metagraywacke) from the *Idzhek Group* yielded the values $T_{Nd}(DM)$ = 2.4 and $T_{Nd}(DM\text{-}2st)$ = 2.5 Ga (Table 4, Fig. 3). This data set demonstrates again that the source rocks of supracrustal Kholbolokh, Kyurikan, and Idzhek groups were of the Early Proterozoic and, to a lesser extent, the Archean age.

Rock samples of the *Fedorov Group* revealed two intervals of model Nd ages. The interval of $T_{Nd}(DM)$ = 3.1–3.2 Ga (Table 4, Fig. 3) characterizes crystalline schists of the hornblende–clinopyroxene–plagioclase type (metabasalts) displaying a rather high level of $^{147}Sm/^{144}Nd$ = 0.1444–0.1599. In their geochemical parameters, these metamorphic rocks correspond to continental basalts. Protoliths of another group of hornblende–diopside–hypersthene–plagioclase schists and biotite ± amphibole–hypersthene plagiogneisses and gneisses are attributed to subalkaline volcanic rocks of the basalt–andesite–dacite–rhyolite series of the island-arc type. In this group, $T_{Nd}(DM)$ values range from 2.2 to 2.4 Ga (Table 4, Fig. 3). The value of $T_{Nd}(DM)$ = $T_{Nd}(DM\text{-}2st)$ = 2.2 Ga is obtained for the Fedorov Group Sample B-2880/1 of hypersthene–hornblende plagiogneiss considered as a petrochemical equivalent of graywacke (Tables 2–4). There is no geological data suggesting any age difference between the two groups of metavolcanic rocks from the Fedorov Group, and we believe that they all were formed in the Early Proterozoic time (~2.2–2.0 Ga ago), as was concluded before (Kovach *et al.*, 1996a). It is likely that volcanic protoliths of the continental and island-arc types originated from mantle sources different in isotopic composition.

DISCUSSION

The new results of Sm–Nd isotopic study show that the high-grade metamorphic supracrustal rocks of the central Aldan Shield were formed in many cases after protoliths, whose material was derived from provenances composed of the rocks with the Late Archean and Early Proterozoic Nd model ages (Fig. 3). The Early Archean model age of source areas can be positively suggested only for two rock types: for the high-Al gneiss of the Kurumkan Group from the Nimnyr block of the AGGT and for the garnet–biotite plagiogneiss of the Kurulta Group from the Olomokit block of the OGGT (Table 1), but this does not exclude a possibility that both rocks were formed at some later stages of geological history of the Aldan Shield.

The Sm–Nd isotopic data obtained leave no doubt that the Chuga, Kholbolokh, Kyurikan, Idzhek, and Fedorov groups of supracrustal AGGT rocks are Early Proterozoic in age, although clastic material derived from the Archean rocks apparently was a component of their protoliths. Isotopic and geochronological data available at present allow us to precisely define the age limits for the Fedorov Group only. Rocks of this group enclose the synmetamorphic intrusions of subalkaline gabbro, diorite, and quartz diorite, which are 1993 ± 1 Ma old (Kotov *et al.*, 1995). This defines the age of the earliest structural and metamorphic transformations of the granulite facies in the rocks under condition. Judging from Nd model ages, the lower age limit for their protoliths is about 2.2 Ga, and the Fedorov Group as whole could have been accumulated within the interval of ~2.2–2.0 Ga.

It is more difficult to estimate ages of the Kurumkan, Amedichi, Zverev, Kurulta, and Seim groups of the AGGT and adjacent OGGT. Some aspects of this problem were considered before (Kovach *et al.*, 1996b), and we can assume now that metapelites of the Kurumkan Group in the western part of the Nimnyr block accumulated above the Early Archean continental crust

($T_{Nd}(DM)$ = 3.5–3.8 Ga; Sal'nikova *et al.*, 1996), whereas in the eastern part of the block (Neelov and Mil'kevich, 1979) their protoliths included clastic materials derived from both the Early Archean continental crust and Early Proterozoic volcanic province (Fedorov Group). Afterward, there were events of tectonic conjoining of the Kurumkan metapelites, infracrustal tonalite–trondhjemites, and Fedorov volcanics, and of their subsequent deformation and metamorphism. This scenario explains why the Kurumkan Group section incorporates rocks, whose model Nd ages range from the Early and Late Archean to Proterozoic, and why age values of this kind for the high-Al gneisses are independent of the geographic position of sampling sites. According to this scenario, the formation period of the Kurumkan group can be approximately restricted by the time interval of 2.2–2.0 Ga and attributed to the Early Proterozoic stage of the Aldan Shield evolution.

However, typical metagraywackes are missing from the Fedorov Group, and this fact contradicts the above scenario. According to new Sm–Nd isotopic data, the garnet ± hypersthene–biotite plagiogneisses from the Kurumkan Group stratotype are similar to biotite and garnet–biotite plagiogneisses of the Chuga Group exposed not far away[5] (Fig. 1). This similarity can be interpreted either as a result of tectonic conjoining of the Kurumkan and Chuga groups in the northwest of the Nimnyr block, or as a consequence of the Early Proterozoic crust-forming processes in the OGGT–AGGT junction zone. At the same time, we cannot exclude a possibility that the Early Proterozoic provenances of rocks similar in composition to the Chuga and Fedorov rock groups contributed their material to metapelites of the Kurumkan Group. Simultaneously, the high-Al gneisses of the group may incorporate components derived from the juvenile continental crust of the Late Archean age (Kovach *et al.*, 1996b). Such a stage of the AGGT geological developments is also inferable form isotopic data for supracrustal rocks and granitoids of the Seim and Sutam blocks.

In contrast to the Kurumkan Group, the Nd model ages for metapelites and metagraywackes of the Amedichi Group suggest that their protoliths included material derived from provenances of the Late Archean mean Nd model age (≤3.0 Ga). It is equally possible that original Lower Proterozoic sediments of this group in addition incorporated the clastic components transported from mixed source areas composed of crustal rocks of the Early Archean and Early Proterozoic ages.

The highly variable model ages of metagraywackes from the Seim group (3.0–2.6 Ga) tolerate the double interpretation: accumulation of original graywackes is related to erosion of either the Archean and Lower Proterozoic rocks in combination, or the Late Archean crust only.

[5] Note that Mironyuk (1961) originally discriminated the Chuga Formation from the Upper Aldan Group of the Iengr Complex.

Consequently, supracrustal rocks of this group are of the Late Archean (not older than 2.6–2.7 Ga) and/or Early Proterozoic ages (no younger than 1.95 Ga). Concrete model ages for the metagraywackes under consideration are comparable with $T_{Nd}(DM\text{-}2st)$ = 3.0–2.8 Ga calculated for the Early Proterozoic granitoids of the Seim block, whose intrusion occurred prior to formation of the Timpton thrust (Kotov *et al.*, 1995), and with $T_{Nd}(DM\text{-}2st)$ values inferred for the tonalite gneiss of the nonstratified infracrustal complex (3.0 Ga), or for Early Proterozoic granitoids of the Sutam block (3.6–3.2 and 2.9–2.8 Ga; Shemyakin *et al.*, 1995). These dates suggest a wide manifestation of the Late Archean crust-forming processes in both AGGT blocks and the same age for the Seim Group.

Source areas incorporating rocks with the Late Archean model ages contributed their material to metagraywackes of the Kurulta and Zverev groups of the Kurulta block, where their protoliths might be accumulated in this or later geological time. Bibikova *et al.* (1989) estimated, however, that the earliest episode of granulite metamorphism in basic crystalline schists (metagabbroids or metabasalts) of the Kurulta Group took place 3460 ± 16 Ma ago. According to Nemchin (personal communication), the schists containing zircons analyzed by Bibikova and her colleagues yielded the Nd model age of about 3.4 Ga and revealed the abnormally low ratio of $^{147}Sm/^{144}Nd$ = 0.1254 as compared to the range of 0.1905–0.2077 characterizing seven other schist samples from the Kurulta Group. This may indicate that the rock in question was contaminated with crustal material and incorporated xenogenous zircon grains. In the light of these results, we cannot argue for sure that the Kurulta Group includes rocks of the Early Archean age.

Considering jointly our and other available data on the Nd model ages for high-grade metamorphic supracrustal and granitoid rocks of the central Aldan Shield (our Table 1; Shemyakin *et al.*, 1995; Sal'nikova *et al.*, 1996), we may recognize the Early Archean (~3.8–3.2 Ga), Late Archean (~3.1–2.8, maybe up to 2.6 Ga), and Early Proterozoic (~2.5–2.0 Ga) stages of continental crust formation in this region.

The Early Archean stage was responsible for the continental crust formation in the northwestern Nimnyr (3.8–3.5 Ga) and Sutam (3.6–3.2 Ga) blocks of the AGGT, and also in the Kurulta and Olomokit blocks (3.5–3.4 Ga) and eastern part of the OGGT (3.7–3.3 Ga). Manifestations of the Late Archean crust-forming events are detected in the eastern OGGT (3.2–29 Ga) and in the Kurulta (3.0 Ga), Zverev (2.9 Ga), Seim (3.0–2.6 Ga), Sutam (2.9–2.8 Ga), probably Nimnyr (3.0–2.9 Ga), and Melemken (3.1 Ga) blocks of the AGGT. The Early Proterozoic age of such events is established in the eastern part of the Nimnyr block (2.4–2.0 Ga) and Fedorov allochthon (2.4–2.2 Ga) of the AGGT. In addition, supracrustal rocks related in origin to reworking of the Early Proterozoic continental

crust are characteristic of the Kholbolokh (2.4–2.2 Ga), Sunnagin (2.6 Ga), Gonam (2.3 Ga), Tyrkan (2.1 Ga), and western Nimnyr (2.5–2.3 Ga) blocks and the Idzhek allochthon (2.4 Ga) of the AGGT. In general, we can see a certain zoning in distribution of Nd model ages: the Olomokit and Kurulta blocks of the eastern OGGT, together with Zverev, Chuga, Nimnyr, Melemken, Seim, and Sutam blocks of the western and central AGGT, display the Early and Late Archean model ages, whereas the Early Proterozoic age values are characteristic of the Kholbolokh, Sunnagin, Gonam, and Tyrkan blocks of the eastern AGGT. There are also the rock zones marked by Early Proterozoic model ages and separating the eastern OGGT from western AGGT (Chuga Group), and also the Nimnyr and Melemken blocks from the Seim and Sutam blocks (Fedorov Group).

Thus, the continental crust formation in central Aldan Shield progressed in a form of discrete relatively short episodes and affected individual structures of the region.

CONCLUSION

New Sm–Nd isotopic data and previous geochronological results (Bibikova *et al.*, 1986; *Drevneishie ...*, 1989; Kotov *et al.*, 1993; Kotov *et al.*, 1995) show that supracrustal rocks of the AGGT and eastern OGGT were mainly formed at the Late Archean and/or Early Proterozoic stages of geological history of the Aldan Shield as it was originally concluded a long time ago (Tugarinov *et al.*, 1967). These data do not exclude, however, a chance that the Lower Archean sedimentary and volcanic rocks are also the constituents of some tectonic blocks of the region.

In the light of new data, we should revise the popular idea that the supracrustal rocks displaying the high grade of metamorphism and widespread in the shield are older than the less metamorphosed supracrustal rocks of greenstone belts located inside the supra- and infracrustal areas. The related idea that the AGGT represents a crustal region of the Early Archean stabilization, where the Late Archean and Early Proterozoic events of activation were manifested only in local narrow zones, must be also revised.

ACKNOWLEDGMENTS

We are grateful to L.M. Bogomolova, G.M. Drugova, V.L. Dook, A.N. Zedgenizov, V.I. Kitsul, S.S. Rozhin, and V.S. Shkodzinskii for rock samples kindly submitted to our investigation. This work was supported by the Russian Foundation for Basic Research, project nos. 95-05-15105 and 96-05-64778.

Reviewers O.M. Rozen and D.Z. Zhuravlev

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The Late Riphean Miroedikha Microbiota from Siberia. Communication 2: Interpretation in Terms of Biotic Paleosuccession

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Received December 11, 1996

Abstract—Changes in the abundance and taxonomic composition of organic-walled microfossil assemblages of the Late Riphean Miroedikha microbiota, one of most diverse and thriving in the Proterozoic, were analyzed at ten discrete stratigraphic levels. The revealed time succession of these assemblages can be considered as an analogue of Phanerozoic coenogenic paleosuccessions characterizing the evolutionary trend from pioneer to postpioneer and climax communities. Some peculiar features of Riphean paleosuccessions and difference in diversity dynamics of morphotypes characteristic of microbiota at the transgressive and regressive stages of sedimentation are outlined. It is noted that consecutive taxonomic changes at these stages are comparable with coenogenic paleosuccessions of the migration and mutation types, and these transformations in microbiotas of the Riphean time apparently were related to the first-order evolutionary cycles of paleobasins. The suggested synoptic model of the Miroedikha microbiota "ontogeny" is important for understanding the progressive complication of the Precambrian microbes.

Key words: *microfossils, Proterozoic, Riphean, sedimentary basins, paleosuccessions, ecosystems.*

Understanding of fundamental evolutionary trends of the Proterozoic biosphere is a keen problem of Precambrian paleontology. This problem cannot be solved by using only traditional methods of actualism for interpretation of the composition and environments of microbiotas discrete in space and time. It requires a new approach; the analysis of discrete microfossil assemblages should be combined with due consideration for their representative regional successions and a detailed examination of dynamic changes in the composition of past communities of microorganisms representing each assemblage (Veis and Vorob'eva, 1992, 1993; Hofmann and Jackson, 1994; Zhang and Walter, 1994; Veis and Petrov, 1994; Burzin, 1996; Petrov and Veis, 1995; Veis *et al.*, 1998a).

Analyzing the Late Riphean Miroedikha microbiota unique in terms of abundance and diversity, we demonstrated in our first communication (Veis *et al.*, 1998b) how environmental factors controlled the facies-ecological distribution of different assemblages and particular morphotypes in the paleobasin. In this communication, we consider processes which were responsible for the regular secular alternation of these assemblages and morphotypes. Our purpose was to show that in the Precambrian, like at the later stages of geological history, these processes developed in a strict chronological order, and this ordering reflected not only the environmental control over distribution of microorganisms and their communities, but also a certain autonomy of microbiota with respect to external factors.

PRECAMBRIAN MICROBIOTAS AND ENVIRONMENTS: STATISTIC AND DYNAMIC ANALYSIS

Initially, relations between the Precambrian organisms and their surroundings, as well as influence of environmental factors on microbiota composition, were examined and interpreted from the standpoint of actualism and actual models. When applied to local settings of shallow Proterozoic carbonate platforms populated by specific microorganisms, this approach was especially successful (Hofmann, 1976; Knoll and Golubic, 1979; Knoll, 1982, 1984; and references therein). More general judgments on lateral taxonomic variations of biotas in different zones of Precambrian basins appeared later (Vidal and Knoll, 1983; Knoll *et al.*, 1991; Vidal and Nystuen, 1990; Zhang and Walter, 1992; Hofmann and Jackson, 1994), but these appraisals were predominantly based on the neontological data and almost ignored a great body of information about the ecological niches of Paleozoic and Mesozoic palynomorphs resembling their ancient analogues (for review and references, see Richardson, 1984; Volkova, 1997). Finally, a few years ago, we got a chance to depict persistent distribution patterns of Riphean biota in the totality of colonized settings in major basin types of that time (Veis and Petrov, 1994; Petrov and Veis, 1995). Principal in the biota structure turned out to be four facial-ecological groupings of microorganisms (deep-water, medium-deep or "optimal," transitional,

and shallow-water) different in taxonomic composition and regularly replacing one another along the basin profile. The main factor that controlled this replacement was the basin depth, whereas the intensity of hydrodynamics and biota relationships with certain stages of sedimentary megacycles acted as additional indirect factors.

It is essential that all aforementioned considerations were basically aimed to analyze the influence of environmental factors on the microbiota composition at the particular stratigraphic levels (statistic approach), and much less attention was paid to such an influence during the representative chronological periods (dynamic approach). As a rule, the investigations resulted in statements characterizing general paleoecological distinctions of one or several model microbiotas from particular Proterozoic time units and in a broad extrapolation of their characteristics over the considerable stratigraphic intervals (*The Earth's*, 1983; *The Proterozoic*, 1992; Knoll and Sergeev, 1995; and other works). Inattention to the chronological aspect of Precambrian paleoecology was quite reasonable in the early time of microphytology. Moreover, exactly this approach finally created the basis that was used to integrate the mosaic of facial-ecological groupings of ancient microorganisms characterizing different chronological levels into a general model of Precambrian biota distribution in space.

Afterward, parallel to accumulation of data illustrating the environmental influence on composition and distribution of microbiotas in Proterozoic basins, the scientific community realized that there is the growing deficit of information concerning the influence of directional secular environmental changes not only on the ecological specifics of particular paleocommunities, but also on the evolutionary transformations in the whole ancient world of microorganisms. It became clear just recently what kind of relationships exist between the reversible changes in Riphean microbiotas and sequential stages (transgressive, regressive, stable) of the basin evolution, and between the irreversible biotic changes and periodical transformations of the basin, which distorted the stage sequence (Veis *et al.*, 1998a). These data revealed the necessity to concentrate attention on the events responsible for significant regional rearrangements in the basins (on "external" boundaries of megacycles) and on concurrent transformations in microbiotas. Our study also showed that precisely these regional or wider paleogeographic changes served as basic factors inducing the periodic destabilization of ecosystems in Precambrian basins. This destabilization, comparable in significance with Phanerozoic biotic crises, gave rise to deep changes in the lateral structure of microbiotas and finally resulted in the taxonomic renewal of highly persistent Proterozoic communities of microorganisms (Veis *et al.*, 1998a).

Analysis of most representative microbiotas at the relatively stable ("coherent" after Krasilov, 1986) periods of their existence is especially promising and allows one to get insight into mechanism of such renewals. Exactly this analysis reveals the successive appearance of distinctive morphotypes (new-comers inclusive) and microfossil assemblages in terms of paleoecology and biotic evolution, thus opening a way to understand an integral model of "ontogeny" of Precambrian ecosystems characterizing complete megacycles. Such a model would be an important step toward the better understanding of progressive complication of the Precambrian organic world.

The approach like this or its particular elements have been used over many years in studies on Quaternary (recent) and Mesozoic communities of marine and terrestrial organisms. The studies revealed that the long-ranged ("polychronous") faunas and floras incorporate some specific taxonomic assemblages replacing one another with time in a certain manner. Series of these consecutive assemblages are termed as successions and paleosuccessions. Their origin and consecutive alternation are under control of both the inner processes in ecosystems and the periodic ecosystem crises in response to the action of external factors (e.g., Krasilov, 1969, 1986, 1997; Kalandadze and Rautian, 1993).

Successions detectable in recent (Quaternary) communities are studied now in detail (for review and references, see *Ekologiya*, 1986). The succession analysis is performed at the population (specific) level that is *a priori* inaccessible in studies on paleosuccessions, when we deal with occasional data samples characterizing ancient taxa. Chronological and time-resolution accuracy feasible in the first and second situation are also incomparable, and, what is especially important, successions and paleosuccessions are under control of different factors influencing their origin and evolution. In contrast to the former, primarily dependent on development of biotic components in ecosystems, paleosuccessions experienced a complex action of diverse processes (Krasilov, 1969) related either to the biota migration and evolution (coenogenic paleosuccessions), or to the burial and preservation of fossil faunas and floras (taphonomic paleosuccessions). Despite the clear difference between successions and paleosuccessions, we may distinguish in both cases certain identical (pioneer, postpioneer or intermediate, climax, etc.) stages of their development. In this work, we use the concept of coenogenic successions most completely developed by Krasilov (1969, 1986, 1972, and references therein) and focus our attention on them leaving aside the taphonomic aspects of studies on Precambrian biotas.

When coenogenic paleosuccessions are under consideration, the most important is the initial (pioneer, aromorphous, macromutation) relatively short stage of their evolution. This stage was preceded by essential events of extinction of many phylogenetic lines and by considerable reduction of populations and their taxonomic diversity. This stage of incoherent ecosystem

state is characterized by isolation of the high-rank taxa, whose appearance was decisive in the further biota specialization for a long period of time. At the next postpioneer and final climax stages, which were longer, faunas and floras demonstrated the growing stability of development, and their diversity progressively increased till the moment when the fragmentation feasibility of ecological niches was exhausted, and afterward this trend slowed a little under the influence of stabilizing selection growing with time. Depending on the procreation rate of unfavorable environments and/or on the progressing activation of inner destructive processes able to eliminate the whole ecosystem, the final stage (climax) may be either present or missing, and this controlled the character of transitional intervals separating the neighboring paleosuccessions (Sukachev, 1928, 1964; Krasilov, 1969, 1986; Kalandadze and Rautian, 1993; Zherikhin, 1997).

When the climax stage got out of the row and biota retreated back into one of the postpioneer states, ancient organic communities demonstrated a sudden speciation outburst in response to the stepwise drop in influence of the stabilizing selection that was associated, in particular, with appearance of unusual or even freakish taxa (monstrous morphotypes). Many scientists are of the opinion that main factors responsible for changes in coenogenic successions are not the external (exogenic) but internal, "endogenic" after Sukachev (1958) or "autogenic" after Krasilov (1969), driving mechanisms of biota evolution. However, the proven indications of this effect are rare, in paleontological records, if at all present, and observable changes in faunas and floras are interpreted as a rule in terms of migratory paleosuccessions.

There were only a few attempts to consider the pre-Quaternary fossil remains from the standpoint of paleosuccession analysis, all devoted to terrestrial plant communities (Krasilov, 1969, 1986, 1997; Meyen, 1988, 1990; *Sovremennaya* ..., 1989; Zherikhin, 1997). Applicability of this approach is thought to be problematic with respect to Phanerozoic marine organisms, and, as far as it concerns Precambrian objects, it has never been discussed either. Moreover, studies on the Precambrian ecology often followed an antipodal procedure of conventional synchronization of fossil communities somehow different in age (Knoll *et al.*, 1991; Petrov and Veis, 1995). This procedure allowed one to decrease or even to undo the negative effect of data incomplete with respect to the facial dependence of microbiotas from different stratigraphic levels and to elucidate their general distribution patterns.

The comparative analysis of facial-ecological groupings of Precambrian microorganisms different in taxonomy but correlative in the lateral aspect (statistical analysis) reveals their distinctions but cannot provide an insight into the secular sequence of the gain of these distinctions. Just studying the entire chronological architecture of different microbiotas (dynamic analysis), it is possible to outline its development with time and to understand the conceivable evolutionary significance of principal replacement stages of microfossil assemblages, including those inadequately preserved in paleontological records and interpretable on the basis of indirect data. In the reconstructed lateral structure of microbiotas, these stages may be manifested merely in a form of qualitative and/or quantitative changes in the composition of certain assemblages, whereas in the chronological architecture they reflect the logical result of the influence of concrete evolutionary factors.

The Late Riphean Miroedikha microbiota from the Turukhansk Uplift of Siberia represented by organic-walled microfossils (German, 1990; Veis *et al.*, 1998b) appears to be one of the best for studying the dynamic aspects of Precambrian ecology for the following three reasons. First, abundance of organic-walled microfossils in deposits of the Miroedikha Formation is extremely high: ten different stratigraphic levels with microorganic remains are detected inside the unit about 210 m thick. Second, these microfossils are perfectly preserved and very diverse in taxonomy (70 species of more than 50 genera). Third, the model of Miroedikha basin development is understood quite well (Petrov and Semikhatov, 1997). The main objectives of this work are as follows: (1) to consider the succession of the Miroedikha fossil assemblages as representing an analogue of Phanerozoic paleosuccessions and to outline its distinctive features characteristic of the Precambrian paleosuccessions; (2) to distinguish stages of diversity variations in the Miroedikha microbiota comparable with those of Phanerozoic paleosuccessions; (3) to elucidate the essence of processes, which controlled the biotic changes with time, and to classify the type of Miroedikha paleosuccession; and (4) to depict the general model of "ontogeny" for the studied microbiota.

SEDIMENTOLOGY OF THE MIROEDIKHA BASIN AND SECULAR CHANGES IN ITS MICROBIOTA

Data on lithology, stratigraphy, sedimentology, and microphytology of the Miroedikha Formation are known quite well from recent publications (Petrov and Semikhatov, 1997; Veis and Vorob'eva, 1998, Veis *et al.*, 1998b). Below, they are considered just briefly in the context of this work.

The Miroedikha Formation of the Upper Riphean of the Turukhansk Uplift represents a relatively thin (160–210 m) unit composed of variegated clay, clay–carbonate, carbonate, and stromatolite rocks (Fig. 1). The rocks are exposed in the western part of the region within the asymmetric Miroedikha syncline. On the basis of available isotopic-geochronological and chemostratigraphic data, the unit is attributed to the middle Upper Riphean (Petrov and Semikhatov, 1997), although biostratigraphic constraints suggest its localization in the upper part of the erathem.

The unit divisible into two subformations conformably rests on the underlying grey dolomites of the Shorikha Formation and shows the gradual transition to the overlying Turukhansk Formation predominantly consisting of red stromatolitic dolomites. The lower Miroedikha Subformation (37–47 m) comprises beds of clayey and occasional stromatolitic dolomites intercalated with shale interbeds, whereas the upper one (120–170 m) is composed of thicker alternating units of carbonate and shale beds. Together with the dolomitic Turukhansk Formation, the Miroedikha deposits characterize an intricate transgressive sedimentary cycle (Petrov and Semikhatov, 1997). In this period of basin evolution, the diverse shallow settings of the inner shelf were replaced by wider, the more stable and uniform zones of the distal shelf, and the influx of fine, predominantly clayey siliciclastic material gradually grew at the beginning and decreased afterward.

Abundant and diverse microfossils are recovered from ten discrete stratigraphic level of the Miroedikha Formation: from four levels of the lower subformation and from other six in two lower members of the upper subformation. The overlying members yielded only scarce, poorly preserved microfossils omitted from consideration in this work. The section interval spanning these levels characterizes the transgressive and regressive trends of the basin evolution, which were separated by a short-term event of the sea-level drop followed by a stage of stable deep-water environments reflecting the rapid flooding of the shelf (Fig. 1).

The lower subformation (32–40 m thick), except for two uppermost bed packets (5–7 m), corresponds to the transgressive trend. It illustrates the gradual flooding over the former shoal carbonate platform of the late Shorikha time and replacement of the peritidal settings by environments of the open shelf. A short period of the quick sea-level drop at the end of the accumulation time of this subformation resulted in subaerial exposure of the inner shelf zone and left behind the aforementioned thin packets of beds of stromatolitic, locally brecciated dolomites.

The stable stage of the regressive trend of the Miroedikha basin evolution that succeeded a swift transgression and shelf flooding corresponds to the lower shale member of the upper subformation (40–45 m thick), whereas the regression course proper is recorded in the overlying, predominantly carbonate member (10–11 m thick). These deposits characterize a transition from the uniform deep settings of the distal shelf, which were typical of the whole region and accumulated clayey sediments only, to the differentiated subtidal settings marked by predominant carbonate sedimentation.

In our previous publication (Veis *et al.*, 1998b), we demonstrated that each of ten microphytological levels of the Miroedikha Formation yielded its own assemblage of microfossils, whose taxonomic peculiarities can be explained in terms of the facial-ecological structure of Precambrian biota (Veis and Petrov, 1994; Petrov and Veis, 1995). The assemblages were least diverse in most shoal settings of the upper subtidal zone (first, second, ninth, and tenth levels). Their diversity perceptibly grew in deeper areas of the lower subtidal zone (third, seventh, and eighth levels) and was at the maximum within the proximal shelf (fourth and sixth levels), being abruptly, but selectively decreased in the deep distal-shelf zone (fifth level). The revealed character of the lateral distribution of the Miroedikha microbiota is considered a consequence of different environmental confinement of its taxa tentatively divided into groups of eury-, meso-, and stenobionts (Veis *et al.*, 1998b). The first group was characteristic of all four types of paleosettings distinguished within the Miroedikha shelf; taxa of the second group inhabited settings of two–three types, and stenobionts populated only one type of possible habitats. Phytoplankton remains simple in morphology along with different ecological types of cyanobacteria represented eurybionts; mesobionts were planktonic and benthonic pro- and eucaryotes, and the highly organized forms of the latter exemplified stenobionts. Most characteristic morpho- and ecotypes of the Miroedikha microbiota are pictured in Plates I–III and also in our preceding work (Veis *et al.*, 1998b), where we demonstrated that the particular morphological groups of microfossils, when thoroughly analyzed, each reveal different behavior, in both the facial-ecological and chronological (pertaining to paleosuccessions) aspects, during the transgressive and regressive trends of the basin evolution. Distinctive features of the groups are illustrated in Fig. 1 and briefly discussed below.

At the transgressive trend (first–fourth levels) and in relatively deep environments of the transient period before regression (fifth level), colonial coccoidal microfossils were most conservative. From the first to fourth level, their diversity and, to a lesser extent, dimensions grew insignificantly and deceased a little at the fifth level. The microfossil assemblage from the first level consisted merely of small shapeless *Myxococcoides* forms; later, it also included species of serial *Sphaerocongregus*, compact spheroid aggregates of *Sphaerophycus* and hollow *Coneosphaera* along with thalluses of *Ostiana* (third level); and, finally, broad *Polysphaeroides* forms enclosed in filamentous sheaths and pseudofilamentous *Palaeopleurocapsa* appeared among other morphotypes (fourth level). At the fifth level corresponding to deeper sedimentation environments, the diversity of colonial coccoidal microfossils was reduced, three last taxa disappeared from its composition, and *Sphaerophycus* forms became dominant.

In the same sequence of events, acritarchs and trichomes demonstrated another style of transformations. At the first and second levels, they are represented only by most simple *Leiosphaeridia* and *Oscillatoriopsis* forms. At the third level, where species of both genera are very abundant and extremely large (up to 1200 and 20 μm respectively), they associate with such remarkable taxa as *Chuaria* and *Cucumiforma* genera of acri-

tarchs and *Filiconstrictosus*, *Cephalophytarion*, and *Palaeolyngbya* genera of trichomes. At the fourth level, both groups of microfossils become even more abundant and diverse, simultaneously preserving the same or slightly lesser size ranges. This level yielded the lophostriated, ellipsoidal, and acanthomorphic acritarchs in association with monstrous envelopes and with trichomes and trichome-like forms displaying their peak abundance exactly here. In addition, this assemblage of septate and pseudoseptate filamentous morphotypes included *Botuobia*, *Cyanonema*, *Calypthotrix*, heterocyte trichomes, *Arctacellularia*, *Trachytrichoides*, *Arthrosiphon*, *Cephalonyx*, problematic remains of red algae (bangiophytes), and some other taxa. Acritarchs of the fifth level corresponding to the deepest sedimentation settings are much less abundant, smaller in size, and all their characteristic morphotypes, except for problematic *Spumosina* and occasional *Trachyhystrichosphaera*, are missing here. Abundance of trichomes and trichome-like forms is reduced here even to a greater extent, and their representatives, other than the scarce *Oscillatoriopsis* and *Palaeolyngbya*, become extinct.

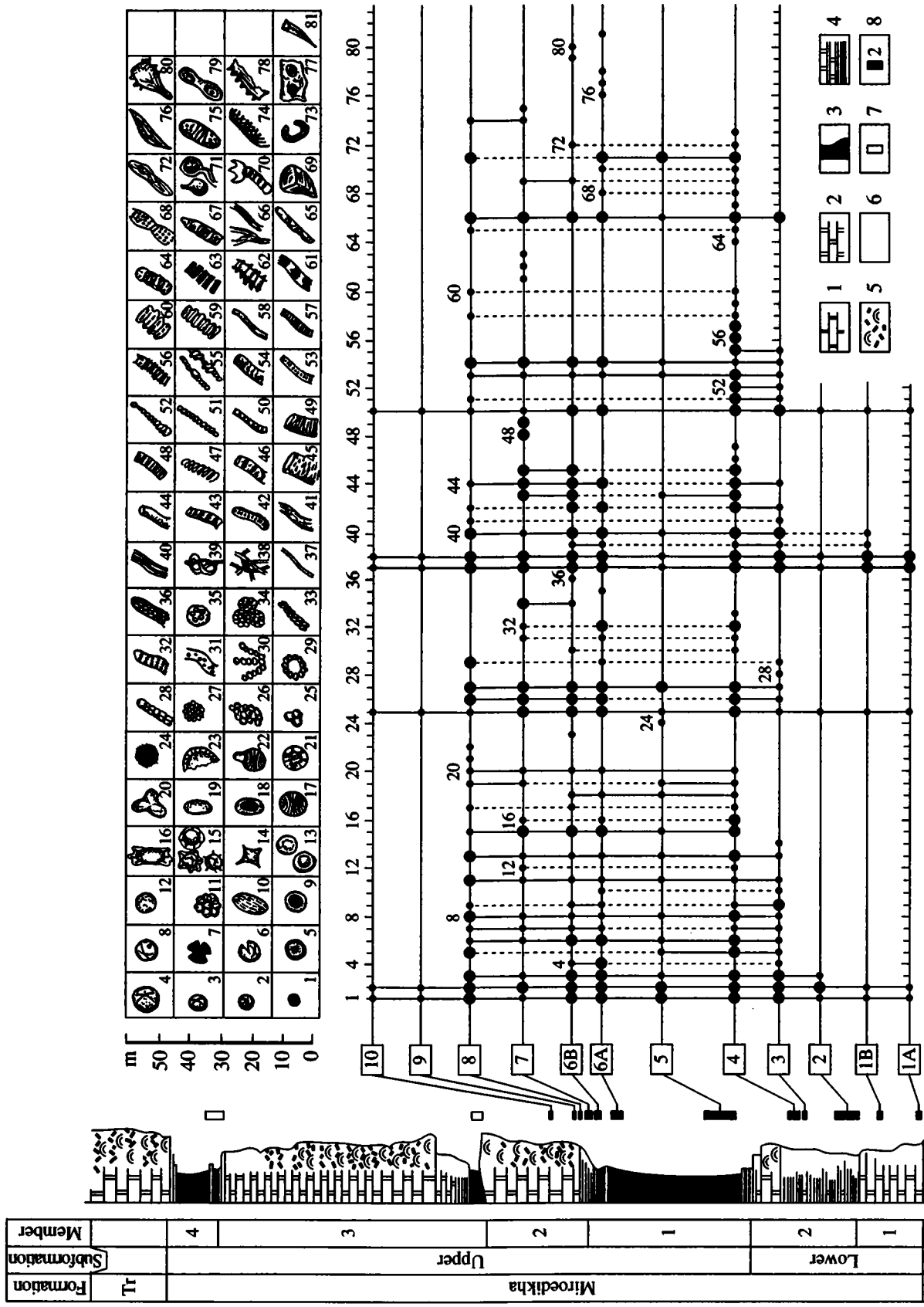
Sheaths characterizing the transgressive trend of the basin evolution resemble in behavior the colonial coccoidal microfossils still being more changeable in dimensions and diversity. In addition to rare *Glomovertella* and *Polytrichoides* recovered from the upper sublevel of the first level and also to simplest *Eomycetopsis* and *Leiotrichoides* dominant at this and second levels, the third one, where both former taxa are extremely abundant, marks the first occurrence of *Siphonophycus* forms displaying the transverse striation and aggregates of very thin *Eomicrocoleus* sheaths. Deep-water deposits of the fifth level yielded a similar assemblage of narrow and medium-sized sheaths, but the aforementioned glomerulate, ornamented, and aggregated forms are missing here, and *Polytrichoides* forms are less abundant. In contrast, all the listed taxa of the first, second, and third levels coexist with abundant broad forms of *Asperatofilum*, *Taenitrichoides*, spiral *Obruchevella*, and wrinkled-costate *Rectia* at the fourth level marking deposits of the moderately deep proximal shelf. The same assemblage characterizing the inner settings of this shelf (the fourth level sections at the Bol'shaya Shorikha River) also includes abundant tubular smooth-walled forms of the cuticular (?) type. The upper size limit of sheaths is 20 μm at the first and second levels; 40 μm at the third and fifth; and 70 to, occasionally, 130 μm at the fourth. The sheaths were especially sensible to changes in the basin depth, and even if this parameter was growing a little in shoal zones of the Miroedikha shelf (the upper sublevel of the first level), the event immediately caused the diversity increase of microorganisms. In addition to isolated specimens of *Eomycetopsis* and *Leiotrichoides* genera characteristic of the lower sublevel, the upper one includes their mats and representatives of *Glomovertella* and *Polytrichoides*. The shoaling event that is recorded at the second level resulted in the disappearance of these mats together with glomerulate and multiserial forms occurring again at the third level of deeper sedimentation environments.

Microfossils of the complex morphology display a peculiar behavior in this part of the section. At the first, second, and third levels, they are represented only by rare thalluses of small *Pseudodendron* forms, whereas the fourth level corresponding to moderately deep sedimentation environments yielded the much more diverse assemblage consisting of large and abundant species of this genus associated with jointed *Eosolena*, plentiful *Germinosphaera*, *Caudosphaera*, and several morphotypes not yet described. The basin deepening at the fifth level left from this assemblage only *Caudosphaera* and small *Pseudodendron* forms, the former possessing the abnormally long processes.

It is remarkable that the transgression intensity most obviously controlled the abundance and diversity of different ecological components of biota exactly during the early Miroedikha time. At the beginning of transgression, for instance, the lateral expansion of eurybionts was notably greater than that of mesobionts, whose abundance increased much later. In contrast, stenobionts of the complex morphology became abundant just at the transgression peak, and, as a result, biota of this terminal interval of transgression is much more diverse than elsewhere in the lower subformation.

The variable behavior of microorganisms from different morphological groups was equally characteristic of the late Miroedikha regression period (levels from fifth to tenth). At the time of transition from uniform deep environments of the fifth level to the moderately deep "optimal" settings of the sixth level, the abundance and diversity of very large acritarchs (acanthomorphic varieties inclusive) and complex morphotypes were unexpectedly high. The dimensions of these microfossils suddenly increased here: acritarchs and filamentous forms, respectively, are more than two (1200 versus 500 μm) and about three times (44–46 versus 15–16 μm) greater than before. Morphological changes in both groups were also spontaneous, and the lower sublevel of the sixth level marks the appearance of abundant acanthomorphic acritarchs *Trachyhystrichosphaera* and *Prolatoforma* in association with aseptate filaments of *Asperatofilum*, representative giant *Pellicularia* and *Fabiformis* of complex morphology, and a set of morphotypes previously unknown. Colonial coccoidal forms were more conservative at this stage of the basin evolution, but even their group is enlarged at the sixth level and includes the subspheroidal colonies of *Coniunctiophycus*, specimens of *Eogloeocapsa* with well-preserved external envelopes, botryoid aggregates of *Palaeopleurocapsa*, filamentous sheaths of *Polysphaeroides*, thalluses of *Ostiana*, and empty colonies of *Coneosphaera*.

Afterward, there was a certain decline in the abundance of characteristic taxa newly appearing in many groups, but acritarchs and notably smooth-walled or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ornamented tubular morphotypes of the cuticular (?) type became more diverse. The acritarch species detectable below associate more with the following remarkable taxa: with rimed *Cymatiosphaeroides* forms possessing the plentiful thin processes and lophostriated *Valeria* morphotypes (upper sublevel of the sixth level); with ornamented *Cerebrosphaera* specimens (seventh level); and with *Valeria* forms missing from underlying deposits and displaying the hypertrophied lophostriation (eighth level). The monstrous forms are present at the sixth, seventh, and eighth levels, and the latter reveals the peak abundance of acritarchs. This peak is not a consequence of appearance of new forms (disregarding the aforementioned unusual forms of *Valeria*) and just marks the abundance maximum of transient medium-sized *Chuaria* (up to 630 μm), *Ch. globosa*, *Trachyhystrichosphaera*, *Satka*, *Simia*, *Navifusa*-like envelopes, and some other forms.

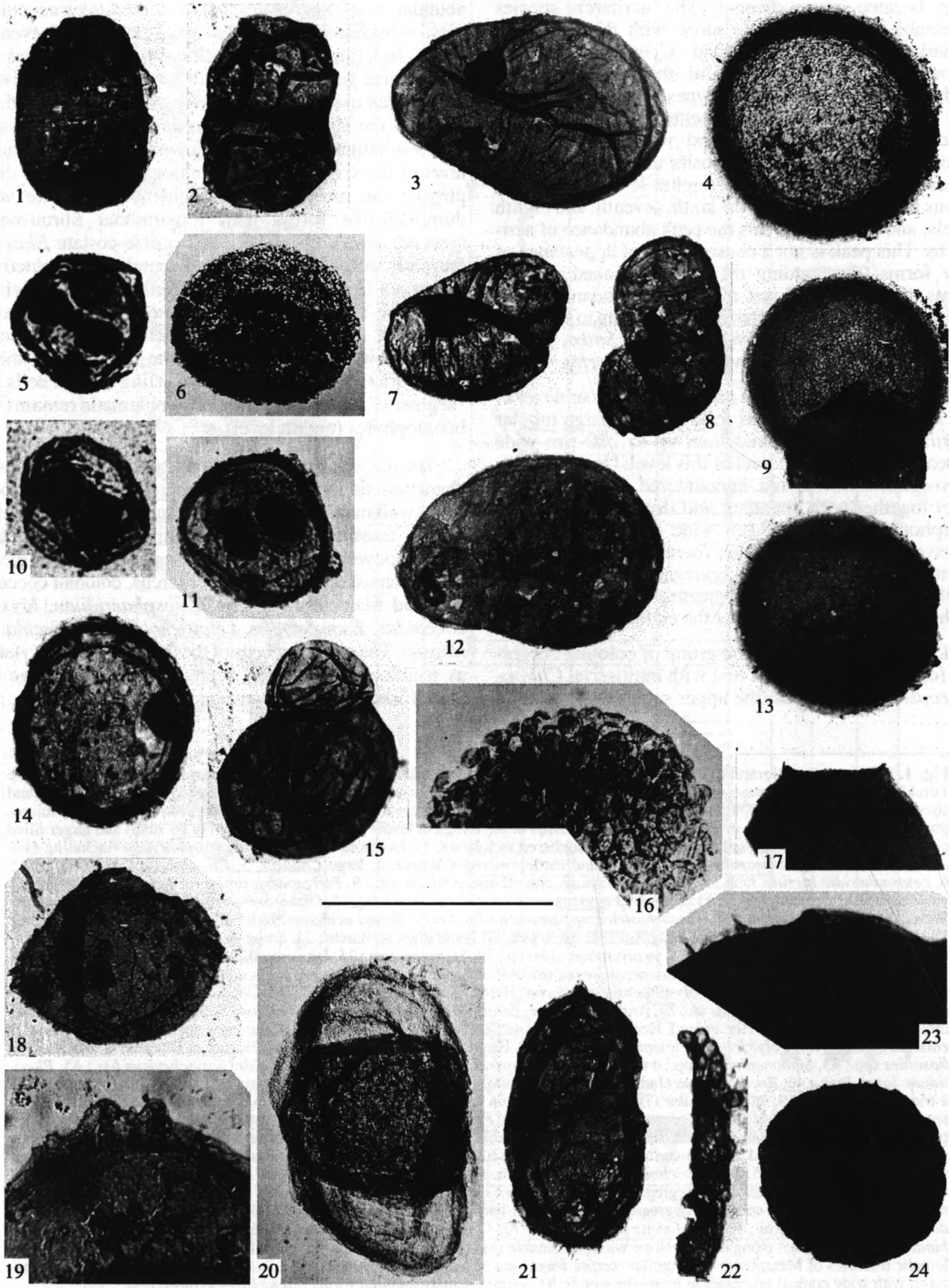
The diversity of sheaths above the lower sublevel of the sixth level increases as follows: very large tubular *Taenitrichoides* and *Plicatidium* up to 260 μm wide appear at the upper sublevel of this level; "banded" *Circumvaginalis* forms are encountered at the seventh level together with abundant and diverse cuticular (?) morphotypes up to 110 μm wide and displaying the transverse structural patterns; *Taenitrichoides* transient from below (82–84 μm) and aggregates of thin usual and fusiform *Eomycmicrocoleus* sheaths, the latter closed at both ends, are characteristic of the eighth level.

During the regression, the group of colonial coccoidal forms was replenished first with multiserial *Chlorogloeopsis* morphotypes (the upper sublevel of the sixth level), and afterwards demonstrated an increase in the abundance of aggregates of *Coniunctiophycus* colonies, subspheroidal *Eogloeocapsa* envelopes (seventh level), and, finally, of empty *Coneosphaera* colonies of several size generations (eighth level). Representatives of complex morphotypes first appeared in this period of time are the elongated-oval *Pseudotawuia* and clavate forms ornamented with wide low spinules (upper sublevel of the sixth level), giant elongated envelopes displaying the problematic multichamber structure and dumbbell-like forms with longitudinal fibrousness (seventh level), spiral fine- and coarse-costate *Elatera* (seventh and eighth levels), and abundant reproductive structures *Germinosphaera* and *Caudosphaera* (eighth level). We should also mention such filamentous septate forms as trichome-like *Trachytrichoides* transient from below (upper sublevel of the sixth level), their large varieties (up to 50 μm) consisting of dark cells or "segments" (seventh level), and problematic remains of bangiophytes (eighth level).

The decreasing diversity and abundance of the biota characteristic in general of the regressive trend is especially well manifested at the ninth and tenth levels, marking the transition to shoal environments of sedimentation. Impoverished microfossil assemblages from these levels consist only of simple acritarchs, colonial coccoidal, and filamentous forms (*Leiosphaeridium*, *Myxococcoides*, *Eomycetopsis*, *Leiotrichoides*, and *Oscillatoriopsis*). The overlying beds of the formation did not yield, as mentioned before, any representative assemblage of microfossils worthy of consideration in this work.

Fig. 1. Scheme of stratigraphic distribution of organic-walled microfossils in the Upper Riphean Miroedikha Formation of the Turukhansk Uplift: (1) limestone; (2) dolomite; (3) shale; (4) alternating shales and clayey carbonates; (5) clastic limestones and stromatolites; (6) pelitomorphic and fine-grained deposits; (7) section intervals sampled in detail and barren of fossils; (8) microphytological levels. Scarcity or abundance of microfossils at phytological levels is indicated respectively by small and larger filled circles. Sketches of microfossil morphotypes are numbered as follows: 1, small and medium-sized *Leiosphaeridia* (including *Leiominuscula*); 2, large *Leiospaheridia*; 3, small and medium-sized *Chuaria*; 4, large *Chuaria*; 5, *Ch. globosa* Ogurtz. et Serg.; 6, *Leiospaheridia bicrura* Jank.; 7, *L. ternata* Tim.; 8, *Nucellosphaeridium* spp.; 9, *Pterospermosimorpha* spp.; 10, *Cucumiforma vanavaria* Mikh. emend. Mikh.; 11, coenobial aggregates of small *Leiosphaeridia*; 12, *Cerebrosphaera* sp.; 13, *Simia nerjenica* A. Weiss; 14, *Octaedrixium* sp.; 15, *Trachyhystrichosphaera* spp.; 16, *Prolatoforma aculeata* Mikh.; 17, *Valeria lophostriata* Jank.; 18, *Konderia elliptica* A. Weiss; 19, *Navifusa*-like envelopes; 20, monstrous acritarchs; 21, *Satka* sp.; 22, envelopes ex gr. *Valeria* with hypertrophied lophostriation; 23, *Cymatiosphaeroides* sp.; 24, *Spumosina* sp.; 25, *Myxococcoides* spp.; 26, *Ostiana microcystis* Herm.; 27, *Sphaerophycus* sp.; 28, *Sphaerocongregus variabilis* Moor.; 29, *Coneosphaera arctica* Hofm.; 30, *Palaeopleurocapsa* sp.; 31, *Entosphaeroides* sp.; 32, *Polysphaeroides filiformis* Herm.; 33, *P. contextus* Herm.; 34, *Coniunctiophycus* spp.; 35, *Eogloeocapsa* sp.; 36, *Chlorogloeopsis* sp.; 37, *Archaeotrichion contortum* Schopf, *Eomycetopsis robusta* Schopf emend. Knoll et Golubic, *Leiotrichoides typicus* Her. emend. Herm., and /or *L. gracilis* Pjat.; 38, mats of sheaths of the same species; 39, *Glomovertella eniseica* (Herm.); 40, *Polytrichoides lineatus* Herm. emend. Herm.; 41, *Eomicrocoleus crassus* Horod. et Donald.; 42, *Archaeoellipsoides* spp.; 43, *Siphonophycus* spp.; 44, *Asperatofilum experatus* (Herm.) and/or *Taenitrichoides jaryschevicus* Ass.; 45, *Plicatidium latum* Jank.; 46, *Rectia costata* (Jank.); 47, *Obruchevella* sp.; 48, narrow and medium-sized cuticular (?) morphotypes with a transverse texture; 49, wide cuticular (?) morphotypes with a transverse texture; 50, *Oscillatoriopsis* spp.; 51, *Filiconstrictosus* spp.; 52, *Cephalophytarion* spp.; 53, narrow and medium-sized *Palaeolyngbya*; 54, *P. sphaerocephala* Herm. et Pyl.; 55, heterocyte trichomes; 56, *Arthrosiphon* spp.; 57, *Botuoba* spp.; 58, *Cyanonema* spp.; 59, *Calyptothrix perfecta* A. Weiss; 60, problematic remains of bangiophytes; 61, *Circumvaginalis elongatus* Serg.; 62, *Rugosoopsis tenius* Tim. et Herm.; 63, large trichome-like forms; 64, *Arctacellularia* spp.; 65, *Trachytrichoides ovalis* Herm.; 66, thalluses of variable size ex gr. *Pseudodendron* with vague longitudinal patterns and their interlacing aggregates; 67, *Cephalonyx sibiricus* A. Weiss; 68, *Eosolena anisocyta* Herm.; 69, rounded-triangular membranes; 70, sublinear aggregates of cells with bud-like terminal structures; 71, *Germinosphaera* spp. and/or *Caudosphaera expansa* Herm.; 72, *Fabiformis porosus* Pjat.; 73, C-shaped remains with longitudinal fibrous structure; 74, *Elatera binata* Herm.; 75, large elongated envelope with presumable multichamber structure; 76, giant form ex gr. *Pellicularia*; 77, problematic thalluses of Metaphyta; 78, lamellar crenate fragments of unknown organisms; 79, *Pseudotawuia* sp.; 80, clavate morphotypes with wide conical processes at their wider ends; 81, narrow acute-triangular forms resembling *Ceratophyton*.

Plate I



Thus, microfossils from the Miroedikha Formation (particularly, meso- and stenobionts of the complex structure) demonstrate an asymmetric type of distribution in the course of the regressive trend of the basin evolution: the maximum abundance and diversity of characteristic transient forms did not persist through the whole period of predominant "optimal" paleoenvironments and were characteristic only of its initial step, when sedimentation environments were deeper than afterward (sixth level). After this diversity peak and until the appearance of shallow settings marked by deposits of the ninth and tenth levels, almost all groups of microfossils show the noticeable diversity decline despite the periodical emergence of certain new forms. It is also important that the abundance and diversity of microfossils from all morphological groups of the Miroedikha biota changed in a contrary manner during the transgressive and regressive stages of the basin evolution, and when one of these parameters grew, another was invariably decreased. In this respect, the considered ecosystems appear to be similar to those of the Phanerozoic and recent time.

DISCUSSION

Turning now to the paleosuccession analysis of the Miroedikha microfossils, we should understand first of all how their successions depended on three main factors, such as the directed environmental changes, biota evolution, and taphonomic transformations of the latter. The relative contribution of these factors in appearance of the recorded microfossil successions characterizing now the Miroedikha Formation from the base upward is of a key significance for a better understanding of the paleosuccession nature.

It is rather difficult to evaluate the role of taphonomic changes in the lifetime structure of a microbiota, because we do not know so far how the Riphean microorganisms were transformed almost immediately into the organic-walled microfossils displaying their cellular structures so perfectly preserved. It is possible just to state that this unique preservation excludes a chance of posthumous transportation of Riphean microorganisms (including those from Miroedikha Formation) far outside of their sufficiently wide facial-ecological niches.

As for two other factors, their particular influence on successions of the Miroedikha microfossils became understandable when we determined positions of all ten distinguished assemblages with respect to the detailed model of the Miroedikha basin evolution (Petrov and Semikhatov, 1997). In this case, we concentrated our attention on the chronological aspect of the problem disregarding the qualitative and quantitative differences between assemblages, which depended on their lateral environmental distribution within the basin. Important in this aspect are the principal trends of the basin evolution, since the short-term events of that or another significance (for instance, the emergence of local marine environments of the early Miroedikha time in the west of the basin, or an episode of regression that separated the accumulation periods of lower and upper subformations) either insufficiently affected the composition of successive Miroedikha

Plate I. Microfossils from the Upper Riphean Miroedikha Formation of the Turukhansk Uplift.

(1–3, 7, 8, 12, 15) envelopes of irregular shape (monstrous acritarchs) ex gr. *Nucellosphaeridium*, specimen nos. 2678-4410, 2678-4406, 2678-4497, 2678-4440, and 2678-4422, (1, 3, 7, 8, 12, respectively) and ex gr. *Leiosphaeridia*, specimen nos. 907-170 (2) and 4084-100 (15).

(4, 9, 13) *Chuaria globosa* Ogurtz. et Serg., medium-sized envelopes with a thick and cancellous peripheral layer, specimen nos. 2578-6844, 2578-6496, and 2578-6480, respectively.

(5, 10, 11) small sphaeromorphic envelopes (? colonial sheaths) with some internal minute bodies possibly representing remains of collapsed daughter cells, specimen nos. 907-410, 907-4120, and 2578-6880, respectively.

(6) *Cerebrosphaera* sp., acritarch with fine wrinkles at the surface resembling those of brains, specimen no. 907–412.

(14, 17, 23, 24) *Trachyhystrichosphaera* sp., envelopes variable in structure and possessing processes: (14) short cylindrical processes embodied in a thin filmy rim, specimen no. 2678-4413; (17) rimless forms with rare wide-conical processes, specimen no. 2578-6430; (23) rimless form with rare narrow-conical processes, specimen no. 2578-6430; (24) magnified fragment of the latter.

(16) *Myxococcoides inornata* Schopf, dense subspherical aggregate of coccoidal cells, specimen no. 907-1810.

(18, 19, 21) *Prolatoforma aculeata* Mikh., ellipsoidal envelopes with short bipolar cylindrical processes enveloped by a filmy frill, specimen no. 2678-4499; (19) higher magnification view of an envelope fragment.

(20) *Pterospermopsimorpha insolita* Tim. emend. Mikh., large thin-walled acritarch with a big inner body, specimen no. 2578-6310.

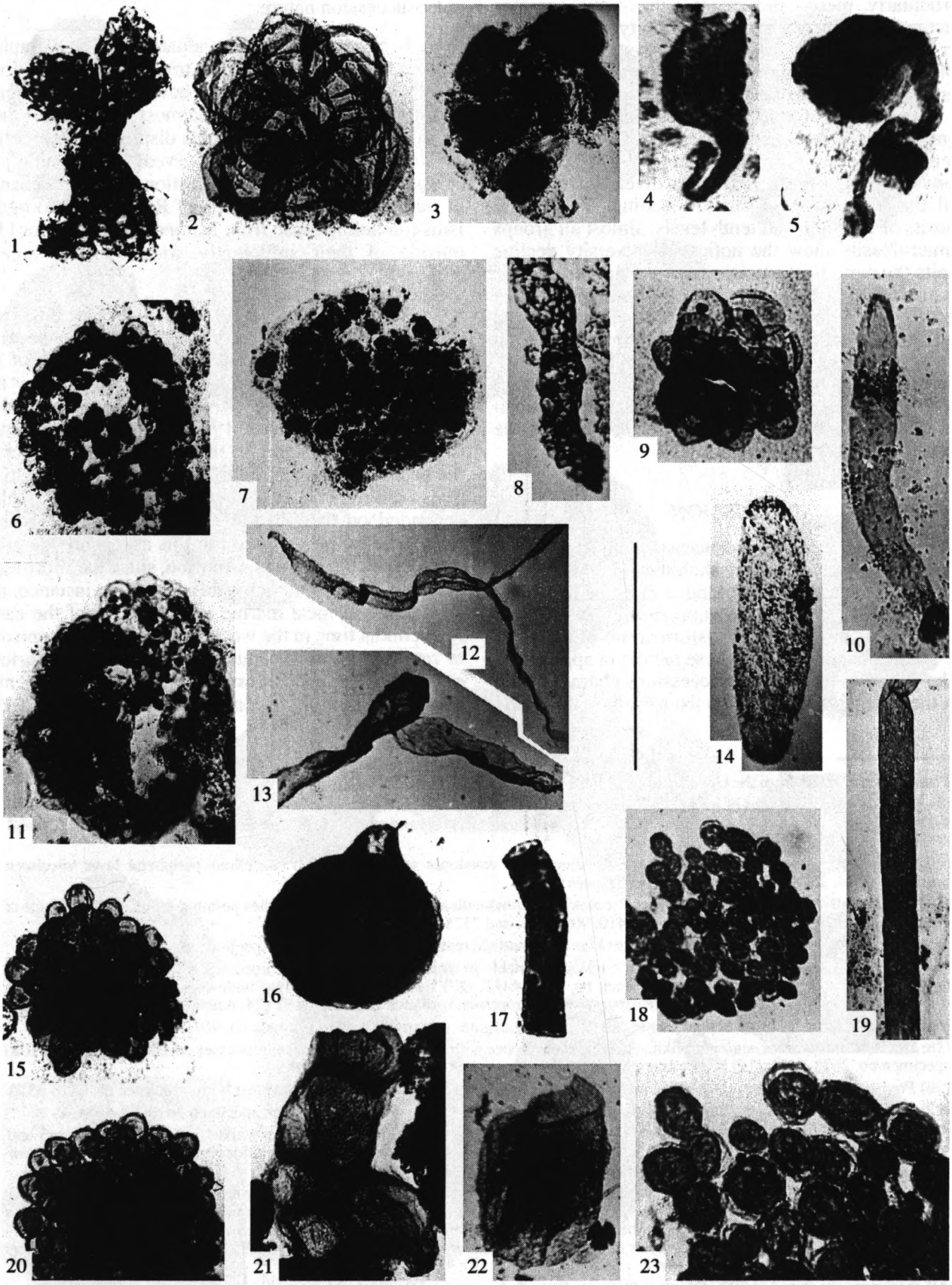
(22) pseudofilamentous aggregate of small coccoidal cells ex gr. *Myxococcoides minor* Schopf, specimen no. 2678-4499.

Scale bar size: 300 μm for images 1–4, 8, 12, 13, 20, 24; 40 μm for image 5; 230 μm for images 6 and 15; 400 μm for images 7 and 9; 55 μm for image 10; 70 μm for image 11; 180 μm for image 14; 110 μm for image 16; 800 μm for images 18 and 21; 700 μm for image 19; 130 μm for image 22; 900 μm for image 23.

The occurrence levels of microfossils: 5, 6, 10 third level, the second member of the lower subformation (Yenisei River); 1, 3, 7, 8, 12, 14, 18, 19, 21, 22, fourth level (ditto); 9, 17, 20, lower sublevel of the sixth level, the first member of the upper subformation (Miroedikha River); 16, upper sublevel of the sixth level, the second member of the upper subformation (Yenisei River); 4, 11, 13, 15, 23, 24, seventh level, the second member of the upper subformation at the Miroedikha River; 2, eighth level, the second member of the upper subformation (Yenisei River).

All specimens are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

Plate II



assemblages, or were not recorded in the available paleontological evidence.

One of most distinct features of the Miroedikha microbiota that is important in the paleosuccession aspect is a different configuration of variation curves depicting the distribution of the biota principal components along the transgressive and regressive trends (Fig. 2). This is most evident in the case of forms attributed to new evolutionary morphotypes, but the same statement is correct, though to a lesser extent, with respect to the interregional taxa and, especially, to members of the pioneer communities (see the respective curves in Fig. 1). The ascending branch of the curves that depicts the behavior of the above microbiota members during the regression is more gentle than that corresponding to the transgressive trend. A certain similarity in structure, composition, and thickness of transgressive and regressive intervals of the formation suggest that the accumulation periods of both parts are comparable in duration. If this is correct, then the character of curves in Fig. 2 may indicate that the time used by the biota to arrive at the peak diversity (i.e., the generation period of "optimal" grouping in its whole range) was longer by transition from shallow to moderately deep settings of the basin than by transition from deep-water environments to the moderately deep ones. The comparative analysis of initial and subsequent states of biota and environments allows us to understand what was behind this distinction.

The early Miroedikha transgression commenced after the late Shorikha stage of the low sea-level stand and predominantly shoal settings in the upper subtidal zone (Petrov and Semikhatov, 1997). According to behavior model of Precambrian biotas in the time of similar extreme events (Veis *et al.*, 1998a), this stage had be marked by degradation of the lateral biota structure and by strongly reduced composition of microorganisms. This is precisely the situation applicable to our case, as one can see from composition of microfossils at the first and second levels marking the initial pulses of the early Miroedikha transgression. These microfossil associations were formed in the upper subtidal zone and consisted of impoverished samples of acritarchs and colonial coccoidal and filamentous forms, except for one assemblage of the upper sublevel of the first level in the westernmost deeper zone of the basin. We noted three distinctive features of these associations. First, they included the most stable and likely highly productive microorganisms, as is evident from the fact of their existence and fossilization in severe shoal settings produced by flooding of the carbonate platform. Second, these organisms were persistent and usually dominant or important members of all

Plate II. Microfossils from the Upper Riphean Miroedikha Formation of the Turukhansk Uplift.

(1) *Ostiana* sp., deformed thallus fragments consisting of compactly arranged coccoidal cells, specimen no. 2678-4222.

(2, 15, 20) coenobial aggregates of small acritarchs *Leiosphaeridia crassa* (Naum.) emend. Jank., specimen no. 2678-4395 (2) and *Nucellosphaeridium minutum* Tim., specimen nos. 2678-4292 (15) and 1427-300 (20).

(3, 7) thin-walled sphaeromorphic envelopes (? colonial sheaths) with rare or frequent internal bodies possibly representing remains of daughter cells, specimen nos. 4075-120 (3) and 2678-4394 (7).

(4, 5) *Caudosphaera expansa* Herm. et Tim., the sprouting structures of ancient fungi or algae in a form of envelopes with thin singular processes, each terminated with a small inflation, specimen nos. 2678-42141 (4) and 2678-42105 (5).

(6, 11) *Coneosphaera* sp., the subspherical empty colonies with a peripheral arrangement of coccoidal cells, specimen nos. 907-1710 (6) and 907-1713 (11).

(8) *Chlorogloeopsis* sp., the filamentous multiserial aggregate of coccoidal cells, specimen no. 907-1320.

(9, 18, 23) *Myxococcoides grandis* Horod. et Donald., the variably shaped aggregates of coccoidal cells at the decomposition stages "punctata," specimen no. 907-460 (9), "capsulata" and "capsulopunctata," specimen no. 2578-6882 (18) and higher magnification view of its fragment (23).

(10) elongated fusiform envelope ex gr. *Brevitrichoides* devoid of its cellular content, specimen no. 2678-42107.

(12, 13) taeniate forms presumably representing ancient green alga *Proterocladus*: narrow dichotomous thallus, specimen no. 2678-6110 (12), and a fragment of medium-sized thallus without dichotomy, specimen no. 907-1820.

(14) *Archaeoellipsoides* sp., the akinete-like ellipsoidal envelope without cellular content, specimen no. 2678-4380.

(16) *Germinosphaera tadasii* A. Weiss, the sphaeromorphic envelope with a wide short process presumably exemplifying the sprouting algal or fungal structure, specimen no. 2578-6420.

(17) *Siphonophycus* sp., a simple sheath with frequent transverse ornamentation, specimen no. 2678-4390.

(19) *Asperatophilum experatus* (Herm.), the structure-less filamentous form (sheath), specimen no. 4075-800.

(21) unusual aggregate of irregular large envelopes (monstrous acritarchs), specimen no. 907-1860.

(22) isolated urceolate envelope (? cell)with a dark shapeless internal body (collapsed intercellular structure), specimen no. 2678-4295.

Scale bar sizes: 260 μm for images 1, 6, 8; 100 μm for images 2, 5, 11, 14; 80 μm for images 3, 7, 9, 10; 150 μm for images 4 and 18; 200 μm for images 12, 13, 22; 400 μm for images 15, 16, 19, 21, 23; 370 μm for image 17; 300 μm for image 20.

The occurrence levels of microfossils: (9) and (20) from third level, the second member of the lower subformation (localities at the Yenisei and Miroedikha rivers, respectively); (1, 2, 4, 5, 7, 10, 14, 15, 17, 22) from fourth level, the second member of the lower subformation (Miroedikha River); (8, 13, 21) from lower sublevel of the sixth level, the first member of the upper subformation (Yenisei River); (18 and 23) from seventh level, the second member of the upper subformation (Miroedikha River); (6 and 11) from eighth level, the second member of the upper subformation (Yenisei River).

All specimens are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

Plate III

Plate III. Microfossils from the Upper Riphean Miroedikha Formation of the Turukhansk Uplift.

(1) *Germinosphaera guttaformis* Mikh., an unusual fusiform envelope with a single, very long process—a reproductive structure of ancient algae or fungi, specimen no. 1407-400.

(2, 7) filamentous microfossil (? sheath) with fine and frequent transverse striation, specimen nos. 2678-4392 (2) and 2678-4205 (7).

(3, 4, 6) *Polytrichoides lineatus* Herm., compact aggregates of subparallel filamentous sheaths of medium size, specimen nos. 4085-400 (3), 2678-4450 (4), and 2678-4309 (6).

(5) fragment of filamentous microfossil displaying a weak longitudinal striation and presumably attributable to ancient scytonemin ean cyanobacteria ex gr. *Pseudodendron*, specimen no. 2678-6495.

(8, 15) *Cephalophytarion variabilis* Schopf et Blacic, trichomes narrowing toward their ends and reduced in width before the terminal cell of a subtriangular shape, specimen nos. 2678-4391 (8) and 2678-4350 (15).

(9, 18, 22) trichome-like remains consisting of cells variable in structure: (9) elongation of cells is insignificant and they are nor constricted near junctions, specimen no. 907-1840; (18) isometric cells are distinctly constricted near junctions, and the form preserved its terminal (?) structure with peculiar corneous processes, specimen no. 2678-4378; (22) form with short and wide cells nonconstricted near junctions, specimen no. 2578-6804.

(10) *Palaeolyngbya sphaerocephala* Herm. et Pyl., the wide curved sheath with trichomes consisting of a series of flat cells resembling a coin stack, specimen no. 907-1801.

(11) *Cephalophytarion turukhanicum* A. Weiss, a trichome constricted at the cell boundaries and possessing narrower ends with specific terminal structures, specimen no. 2678-4381.

(12, 14, 16, 17) *Oscillatoriopsis* sp., fragmented simple trichomes constricted at boundaries between short cells, specimen nos. 2678-4379 (12), 2678-4238 (14), 2678-4203 (16), and 2678-4380 (17).

(13) compact tubular forms (sheaths or possible cuticular morphotypes) with distinct transverse patterns, specimen no. 2578-6842.

(19) *Coniunctiophycus totticus* A. Weiss, aggregates of few or many subspherical colonies of coccoidal cells, specimen no. 907-1350.

(20) *Arctacellularia ellipsoidea* Herm., trichome-like forms representing chains of large urceolate cells, specimen no. 2678-4635.

(21) a fragment of taeniate smooth-walled morphotype (presumable thallus of *Proterocladus* sp.), specimen no. 2578-4130.

Scale bar sizes: 160 μm for image 1; 80 μm for images 2, 9, 18–20, 22; 200 μm for images 3, 4, 10; 235 μm for image 5; 250 μm for image 6; 130 μm for images 7 and 11; 50 μm for images 8, 12–17; 500 μm for image 21.

The occurrence levels of microfossils: (20) from third level, the second member of lower subformation (Miroedikha River); (2, 4, 6–8, 11, 12, 14–18) from fourth level (ditto); (1) from fifth level, the first member of upper subformation (Miroedikha River); (5) from lower sublevel of the sixth level, the first member of the upper subformation (Yenisei River); (9, 10, 19) from upper sublevel of the sixth level, the first member of upper subformation (Yenisei River); (13, 22) from seventh level, second member of the upper subformation (Miroedikha River); (3 and 21) from eighth level (ditto).

All specimens are stored at the Geological Institute, Russian Academy of Sciences, Moscow.

other Miroedikha assemblages from a wide spectrum of settings, thus, representing the group of eurybionts. Third, the microfossils under consideration display very broad time-ranges and are present in most Proterozoic microbiotas different in age. In terms of the paleosuccession analysis, microfossils from the first and second levels may be interpreted as typical representatives of pioneer communities. These communities of archaic type were widespread in the Riphean time and apparently responsible for a swift and wide recolonization of newly formed settings that resulted in biogenic modification of these settings thus prepared for introduction of more complex components of microbiotas sensitive to external factors.

Taxa, which are somewhat less widespread and appeared first in assemblages of the third level (lower subtidal zone), may be attributed to the postpioneer (or late pioneer) communities[1] of the Miroedikha microbiota. These were coenobial aggregates and large isolated *Leiosphaeridia* envelopes, medium-sized *Chuaria*, filamentous *Polytrichoides*, *Eomicrocoleus*, *Glomovertella*, dense mats of simple *Eomycetopsis* and *Leiotrichoides* sheaths, and narrow *Pseudodendron* thalluses. Two of these forms (*Glomovertella* and *Polytrichoides*) are recorded for the first time in the upper part of the first level of the western basin area (Yenisei River). Their appearance at this level was a consequence of occurrence of relatively deep-water environments in this area, and both forms are missing in the second level corresponding to typical shallow-water deposits of more proximal zones of the basin. Other, undoubtedly postpioneer, morphotypes displaying the narrower facial-ecological distribution join the aforementioned forms of indistinct postpioneer affinity at the third and fourth levels. Initially (third level), these were *Pterospermopsimorpha*, *Nucellosphaeridium*, *Ostiana*, *Sphaerophycus*, *Sphaerocongregus*, *Archaeoellipsoides*, *Asperatofilum*, *Filiconstrictosus*, *Cephalophytarion*, *Palaeolyngbya*, *Pseudodendron*, and *Germinosphaera*. Later on, giant *Chuaria*, *Cucumiforma*, *Simia*, *Coneosphaera*, *Arctacellularia*, *Trachytrichoides*, and some other morphotypes joined the taxa listed above, and the diverse assemblage of postpioneer microorganisms was finally formed at a sufficiently mature stage of the transgression development.

At the fourth level (proximal shelf), after appearance of all listed taxa, biota of the early Miroedikha time became saturated with a series of distinctive forms

[1] The successive appearance of postpioneer taxa in the Miroedikha microbiota is described below with due regard for data on their distribution in particular rock samples from respective microphytological levels.

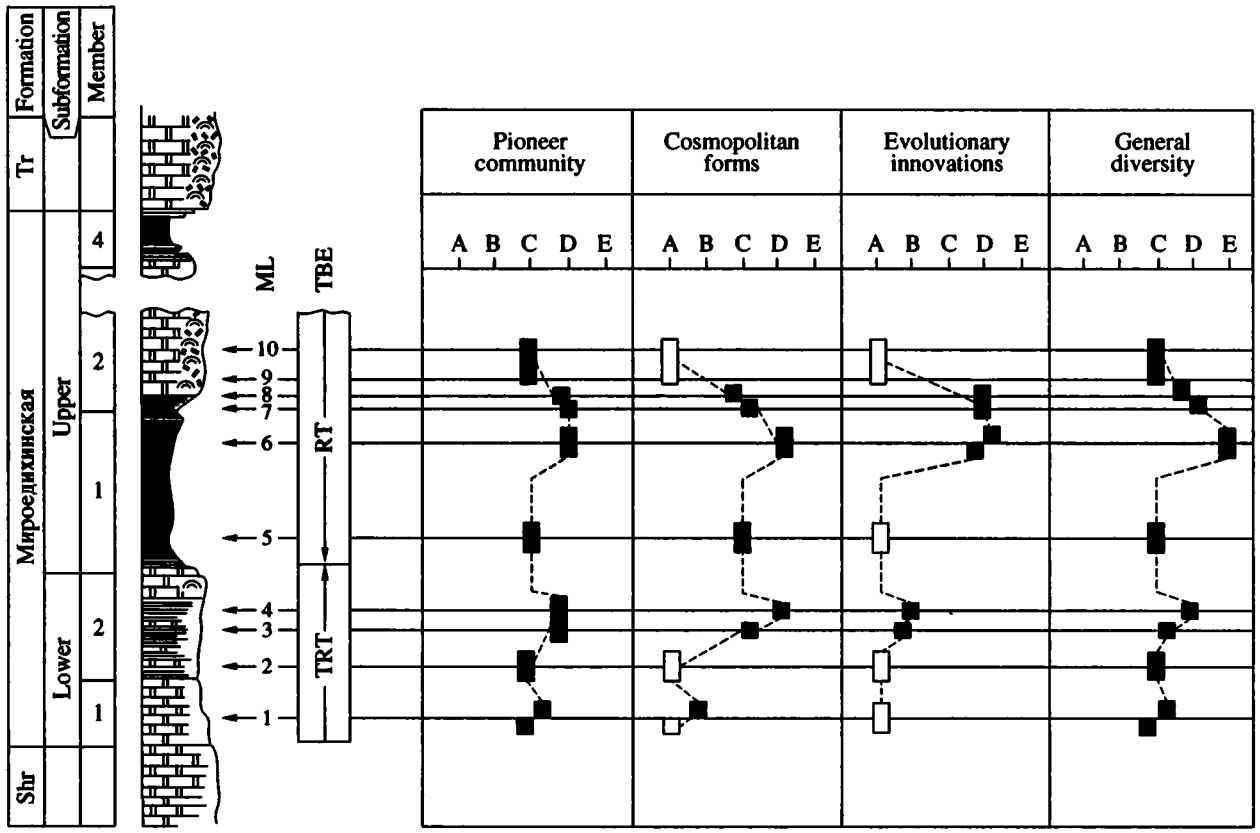


Fig. 2. Abundance of principal components of the Miroedikha paleosuccession and microbiota as a whole in the course of transgressive and regressive stages of the basin evolution: (ML) microphytological levels; (TBE) trends of the basin evolution; (TRT) transgressive trend; (RT) regressive trend. Five conventional abundance gradations are as follows: (A) lack of microfossils; (B) single finds; (C) few dozens of specimens; (D) many dozens of specimens; (E) hundreds of specimens. Data on deep-water grouping are extrapolated to the whole interval of respective environments (other symbols as in Fig. 1).

characteristic of communities of the climax or comparable types. These morphotypes were represented by *Trachyhystrichosphaera*, *Prolatoforma*, *Valeria*, *Polysphaeroides filiformis*, *Taenitrichoides*, *Rectia*, *Arthrosiphon*, *Cephalonyx*, *Eosolena*, *Fabiformis*, *Caudosphaera*, and formerly unknown microfossils. The climax (subclimax) type of communities recorded at this level is inferable from the coexistence of monstrous acritarchs and evolutionary new-comers, *Cephalonyx* and *Arthrosiphon* inclusive, with some very complex microorganisms. Among possible reasons responsible for the appearance of these complex forms (presumably highly organized) in the postpioneer and climax communities, we may indicate two: (1) the true evolutionary innovation (epigenesis) based on transmutation of some (maybe missing from paleontological record) pioneer or postpioneer taxa, and (2) appearance of immigrants from other, formerly isolated areas.

The data presented above allow us to suggest that the impoverished initial (pioneer and postpioneer) and subsequent, more diverse communities of microorganisms were formed basically, or up to a certain time exclusively, due to immigration of very stable transient forms, whose influx progressively grew in the course of transgression development and appearance of favorable habitats. Another process more essential in biotic aspect, namely, the emergence of certain characteristic evolutionary new-comers in the "optimal" biota grouping, was manifested only at the late stage of the early Miroedikha transgression (fourth level) in the taxonomically diverse communities of microorganisms. During this time, the addition of immigrants resulted in a further growth of abundance and diversity of basic detected morphotypes, such as transient acritarchs, colonial coccoidal forms, sheaths, and trichomes, whereas the evolutionary innovations left their imprints amid the marginal acritarchs (monstrous envelopes), trichome-like and forms of the complex structure. In light of these data, the succession of the Miroedikha microfossils characterizing the transgressive trend may be considered quite assuredly as an analogue of coenogenic paleosuccessions of the migration type described by Krasilov (1969).

Behavior of the Miroedikha microbiota by transition from the moderately deep (lower subtidal zone and proximal shelf) to deep (distal shelf) settings is poorly understood because of the inadequate microphytological characteristics of the section interval separating the

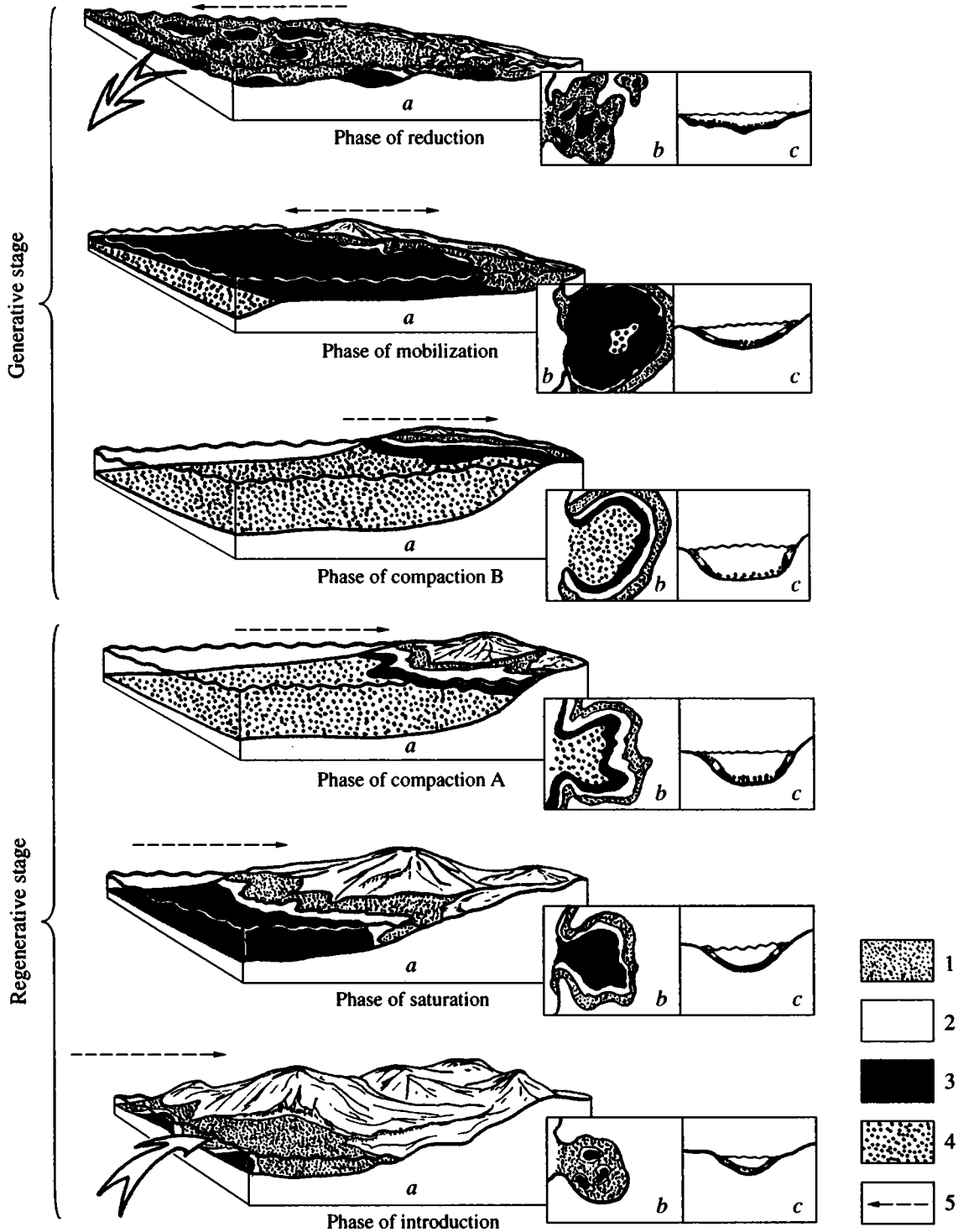


Fig. 3. Model of the Miroedikha biota "ontogeny": (*a*) block-diagrams, (*b*) general plan, and (*c*) profile of the basin; main facial-ecological groupings of microorganisms: (1) shallow-water; (2) transitional; (3) moderately deep-water (optimal); (4) deep-water; (5) dominating direction of a shoreline shift.

fourth and fifth microphytological levels. This situation is a consequence of several concurrent events in the Miroedikha basin history: of short-term shoaling episodes accompanied by a wide development of stromatolites; of partial exsiccation at the end of the early Miroedikha time; and of spontaneous flooding in the region at the very beginning of the late Miroedikha time (Petrov and Semikhatov, 1997). The biotic importance of these events

was, however, insignificant, since all indications of changes in the biota structure at that time are missing from the paleontological record.

In contrast, stable deep-water environments, whose emergence marks the commencement of the late Miroedikha time (fifth level), were of a radical significance for biota development. This event, as we have seen above, fundamentally reduced the taxonomic composition of microbiota, especially of trichomes and complex morphotypes. In the course of the basin deepening, both groups preserved only rare and simple transient components (Fig. 1). The medium-sized *Chuaria*, *Myxococcoides*, *Sphaerophycus*, and *Polytrichoides* existed in this case longer than other members of biota, and problematic *Spumosina* forms represent the only new-comers of the period under consideration.

Judging from published data (Oehler, 1977; Vidal and Nystuen, 1990; Veis and Petrov, 1994), the dramatic diversity decline in all morphological groups of microfossils was quite characteristic of Precambrian deep-water settings and comparable to some extent in its destructive influence with the ecological crises in response to the emergence of shallow-water environments. In contrast with the latter, however, the deep-water stages in the history of Riphean basins never resulted in the complete degradation of the lateral structure of microbiotas and did not destroy the continuity of their development. They just were responsible for migration and/or spatial condensation (contraction) of habitats populated by representatives of the "optimal" and other, more proximal groupings. As for the impoverished deep-water Miroedikha assemblage from the fifth level chiefly consisting of postpioneer taxa, its interpretation in terms of the paleosuccession analysis suggests that this assemblage exemplifies a community of the preclimax (more precisily, disclimax) type.

The continuous succession of deposits characterizing the transition from deep to moderately deep settings of the regressive trend allows us to get further insight into changes in the composition and facial-ecological structure of the Miroedikha biota. At the commencement of this transition, there was a diversity outburst in all morphological groups of microfossils. For instance, their assemblage from the lower sublevel of the sixth level (proximal shelf) includes a series of very indicative forms: giant *Chuaria*, acanthomorphic *Trachyhystrichosphaera* and *Prolatoforma*, oval *Cucumiforma* and *Konderia*, colonial coccoidal *Coneosphaera* and *Polysphaeroides*, ellipsoidal *Archaeoellipsoides*, very large *Fabiformis* of the complex structure, *Eosolena*, and broad *Pseudodendron* thalluses. Almost all listed microfossils occur in many Late Proterozoic microbiotas (Yankauskas *et al.*, 1989; Hofmann and Jackson, 1994; Butterfield *et al.*, 1994) and classed with cosmopolitan taxa, the typical habitants of "optimal" zones in the basins of that time (Veis and Petrov, 1994; Petrov and Veis, 1995). In the Miroedikha assemblages, they coexist with some forms attributable to evolutionary new morphotypes. These are large *Pellicularia* forms with the longitudinal rima, metaphyte fragments, and some morphotypes not yet described, such as remains of narrow conical membranes, sublinear cell aggregates with bud-like termination, and lamellar specimens with aciculate spinules decorating their outer edges.

At the upper sublevel of the sixth level characterizing a transition from the proximal shelf to the lower subtidal zone, i.e., to the typical habitat of the "optimal" grouping, the abundance of immigrants decreased, but the evolutionary new-comers became more frequent. Immigrants from this sublevel are the following forms: acritarchs *Valeria* and *Cymatiosphaeroides*; colonial *Ostiana*, *Chlorogloeopsis*, *Coniunctiophycus*, and *Palaeopleurocapsa*; *Taenitrichoides*, *Plicatidium*, and *Siphonophycus* sheaths, the latter with coarse transverse patterns; and remains of trichome-like *Trachytrichoides*. Complex forms from this sublevel are represented by very broad *Pseudodendron* specimens and by remains of rounded-conical membranes. Evolutionary modifications were most characteristic of acritarchs (their abundant monstrous morphotypes), clavate forms complex in structure and decorated with low conical spinules at the wider end, and elongated ellipsoidal bodies resembling *Pseudotawuia*.

The very high diversity and abundance of both assemblages from the sixth level, the younger of which includes monstrous acritarchs and evolutionary new complex morphotypes not recorded below in the section, allow us to think that the shelf habitats of the late Miroedikha time were populated by assemblages of the climax type immediately after the commencement of regression, and these microorganisms were comparable or even exceeding in diversity the assemblages from the fourth level. In this respect, the obtained data on secular biotic changes recorded at the transition from deep to moderately deep paleoenvironments are consistent with our early model of the lateral biota distribution (Veis and Petrov, 1994). In both cases, we detected the quick appearance of remarkable complex morphotypes in microbiotas. At the same time, the revealed chronological succession of these changes allows us to argue for the accelerated replacement of one assemblage by another as compared with previously described examples. It is likely that this acceleration of biotic transformations in the course of regression is a consequence of a limited migration of the medium-deep settings populated by the completed "optimal" grouping of microorganisms.

Microfossil assemblages from the seventh and eighth levels characterizing the domination of settings of the lower subtidal zone illustrate the microbiota behavior in the course of subsequent development of regression. Microfossils of these levels are still quite diverse, but this diversity is not as impressive as that recorded at the transition from deep to moderately deep

environments (Fig. 1). The depletion in taxonomic composition of the assemblages from the seventh and eighth levels differentiates them from assemblages of the climax type, which are represented, in our opinion, by microorganisms occurring at the fourth and sixth levels. However, the seventh level yielded the representative set of cuticular (?) forms not yet described, and, in addition, we encountered in abundance the monstrous acritarchs and some complex forms interpreted as results of evolutionary innovations and dispersed throughout the seventh and eighth levels. These finds suggest at least the subclimax state of respective communities of microorganisms.

One more important feature of microfossil assemblages from the seventh and eighth levels is the gradual removal of important immigrant taxa from their composition in the course of advancing regression and this removal contrasted with the simultaneous appearance of evolutionary innovations. Most remarkable amid the remaining immigrants were *Trachyhystrichosphaera*, *Prolatoforma*, *Valeria*, *Navifusa*-like envelopes, *Ostiana*, *Ciniunctiophycus*, *Coneosphaera*, *Plicatidium*, *Rugosoopsis*, *Circumvaginalis*, *Trachytrichoides*, *Filiconstrictosus*, *Elatera*, *Caudosphaera*, and problematic remains of bangiophytes. Innovations were represented by the aforementioned smooth-walled and ornamented tubular remains of the cuticular (?) type, by large elongated envelopes displaying a kind of multichamber structure, by *Valeria* specimens showing the hypertrophied lophostriation, and maybe by *Cerebrosphaera* forms.

Thus, the consecutive changes in microbiota at the regressive stage can be interpreted, with certain reservations, as exemplifying the rare mutagenic type of coenogenic paleosuccessions (Krasilov, 1969, 1986). At this stage, the distribution patterns of principal ecotypes (eury-, meso-, and stenobionts) actually were the mirror reflections of the relevant patterns characterizing the transgressive trend. The sole difference between them was the considerably greater abundance of meso- and stenobionts in shallow settings of the regressive stage.

Microfossil assemblages from ninth and tenth levels characterize the shallow settings of the upper subtidal zone, whose emergence documents the regression termination. They include only scarce taxa of the pioneer communities: *Leiosphaeridia*, *Myxococcoides*, *Eomycetopsis*, *Leiotrichoides*, and *Oscillatoriopsis*. In the biotic aspect, this means an abrupt end of the paleosuccession under consideration and commencement of the crisis in evolution of the Miroedikha microbiota in extremely shoal settings.

Thus, the quick recurrent regeneration of biota at the sixth level that is characteristic of the regressive trend is not a continuation of a long series of biotic events responsible for the complete recolonization of the basin in the course of the early Miroedikha transgression. As we have seen, these events included the expansion of pioneer microorganisms, the progressive complication of the postpioneer communities by means of introduction of basic immigrants, and the arrival of these communities at the final climax state of saturation. During the regression, the transition to the final state of these events was secured by the recurrent migration of principal facial-ecological groupings of the Miroedikha microorganisms (the migration apparently was limited in the paleogeographic sense). As a result, immediately after the transition from deep (fifth level) to moderately deep environments (sixth level), the biota reached, consuming its biological resources to a minimum extent, the double effect: (1) it completely regenerated its lateral structure and considerably replenished the taxonomic diversity that was abruptly reduced in deep-water settings; (2) reparation of both parametric indications took a very short time. It seems reasonable to think that the state of stress, when biota was condensed in the deep water settings, created prerequisites for the subsequent development of important evolutionary innovations.

CONCLUSION

As one can see, modifications of the taxonomic composition and facial-ecological structure of the Miroedikha microbiota are divisible in two stages, one of which corresponded to the transgressive and another to the regressive trends of the basin evolution. Biotic changes during the transgression can be classed with coenogenic paleosuccession of the migration type. Initially, the shallow-water ecological niches newly formed in the basin were populated everywhere by most simple transient eurybionts of the pioneer communities, and this ensured the broad recolonization of the niches. Subsequent postpioneer members of the paleosuccession were basically represented by more diverse mesobionts, which demonstrated the progressively growing selectivity with respect to the facial-ecological confinement and predominantly populated the moderately deep or transitional habitats. The paleosuccession terminated with the appearance of a very diverse community of the climax (subclimax) type, where the pioneer and postpioneer taxa coexisted with stenobionts displaying the most complex morphology and distributed almost exclusively in the moderately deep settings of the "optimal" zone.

Due to the prevalence of stable deep-water settings before the regression at the commencement of the late Miroedikha time, the reconstructible part of the basin was occupied by single facial-ecological grouping, whose distinctive features were, as we have seen, poor taxonomic diversity and low abundance of microfossils. In terms of the paleosuccession analysis, this grouping represented the biotic community of the preclimax (or rather, of the disclimax) type.

Microbiotic changes in the course of regression resulted in the relatively quick regeneration of the extreme (climax) state that required the limited recur-

rent migration of the most diverse and abundant "optimal" grouping. An important feature of microbiota behavior at the regressive stage was the highly limited influx of cosmopolitan forms in contrast to the permanent appearance of few, but rather spectacular evolutionary innovations among microorganisms. New taxa controlled the development of biota specifics, and this allows us to compare the revealed series of transformations with the coenogenic paleosuccession of the mutagenic type. Terminal events of the regression created the widespread shallow settings in the basin, sharply reduced the lateral structure and taxonomic composition of the Miroedikha microbiota, and disposed it to the crisis development.

The discussed data on peculiar events in the Miroedikha biota evolution against the background of transgressive and regressive trends of the basin development can be integrated in a synoptic model of its "ontogeny" that may be of a more general significance (Fig. 3). In this model, the attention is primarily focused on the behavior of the "optimal" grouping in space and with time, because it was, in contrast to others, not only the most abundant and diverse, but also the most sensitive to past environmental changes (Veis and Petrov, 1994; Petrov and Veis, 1995; Veis *et al.*, 1998a, 1998b). The analyzed composition and proportions of different morphotypes constituting this grouping in various periods of the Miroedikha basin history allow us to distinguish the regenerative and generative stages of its evolution corresponding, respectively, to the transgressive and regressive trends of the basin development. Each of the stages is divided into three phases: phases of introduction, saturation, and compaction *A* of the regenerative stage; and phases of compaction *B*, mobilization, and reduction of the generative stage.

The function of the regenerative stage (microphytological levels from first to fourth inclusive) was the complete recolonization of all ecological niches within the shallow to deep-water zones of the basin and the development of the "optimal" grouping within the moderately deep zone.

Peculiarities in the composition and lateral structure of microbiota at the beginning of the Miroedikha time suggest, with due regard for the transient character of its taxonomic majority at the transgressive trend, that the "optimal" grouping was initially formed almost exclusively from taxa which migrated from other external basins. As one can see in Figs. 1 and 2, namely immigrants (cosmopolitan taxa) were responsible for the increasing diversity of the grouping at the phases of introduction and saturation of the regenerative stage. These phases were different primarily in the scale and continuity of the "optimal" grouping habitats. At the phase of introduction (first and second levels), its habitats were of the island type, whereas at the phase of saturation (third level) the grouping colonized a vast continuous area (Fig. 3 in this work and Fig. 2 in Veis *et al.*, 1998b).

The natural termination of the saturation stage corresponded to invasion of simple to most complex organisms in the moderately deep zone of the basin; the invasion was adequate to ecological parameters of the basin and resulted in the formation of a stable ecosystem. In the paleosuccession aspect, the invasion produced two effects: (1) the "optimal" grouping formerly ephemeral (at the phase of introduction) advanced at the saturation phase toward the state of progressively more and more diverse abundant assemblages approaching the final climax; (2) it initiated processes of autogenic changes in microbiota. Nonetheless, if these two phases of the regenerative stage were unable to advance the microbiota to the climax possessing the potency of evolutionary modifications, it reached this result only at the phase of compaction *A* (fourth level) terminating the regenerative stage.

At the last phase corresponding to the transgression peak and maximum expansion of deep-water settings populated by respective microorganisms, "optimal" settings recurrently contracted up to a narrow stripe within the proximal shelf and/or lower subtidal zone. The existence period of the "optimal" grouping in the extremely compact state within a linear area at the compaction phase *A* had far reaching consequences. Under conditions of the progressing deficiency of space and resources characteristic of this phase, the separating effect of stabilizing selection was especially powerful and cleared the unstable and nonviable components accumulated before from this representative grouping of microorganisms. The phase of compaction *A* and regenerative stage as a whole terminated with the emergence of widespread deep-water environments (fifth level), when the "optimal" and proximal groupings of microorganisms were forced out of the reconstructible part of the Miroedikha basin.

Autogenic modifications in microbiota apparently were much more intense at the generative stage (levels from sixth to tenth). The first phase of this stage (compaction *B*) resulted in the recurrent regeneration of the lateral structure and climax (subclimax) state of the "optimal" grouping, which migrated to its original position in the basin following the development of moderately deep settings (sixth level). Like at the previous phase of contraction *A*, this grouping occupied a narrow sublinear band and was very abundant and diverse. The steady state and even certain increase of both parameters at the beginning of the generative stage stimulated the evolutionary potency of microbiota that was the main result of that stage.

Transition from the phase of compaction *B* to the phase of mobilization (seventh and eighth levels) was an important event that gave rise to this stimulation. With respect to paleosuccession, the transition signified an abrupt, almost momentary expansion of habitats of the "optimal" grouping thus deprived of its climax state and retreated back to the community of the preclimax type. In Phanerozoic paleosuccessions (Krasilov, 1969,

1986), this type of changes was followed by an outburst of evolutionary modifications in response to the spontaneous intensification of changeability without the selection control. It is remarkable that in the Phanerozoic paleosuccessions these changes were more characteristic of their initial but not terminal phases, though they were recently interpreted in terms of the model of destructive phylocoenogenesis (Zherikhin, 1997). In our case, we may conclude that the phase of mobilization in the Miroedikha biota evolution was peculiar, since the moderately deep zone progressively expanding at that time in the course of regression was invaded by evolutionary new-comers, but not by the transient interregional taxa, as it was characteristic of the regenerative stage.

It is important to stress that our model is in accordance with the theory of free ecospace (Valentine, 1995) shared now by many scientists. According to this theory, the early and apparently quick (in the geological sense) mass origination of principal structural patterns of multicellular Precambrian organisms (maybe Phanerozoic as well) was related not only to critical structural rearrangements in genomes of possible ancestral forms, but also to the existence or periodic appearance of vast and free adaptive zones in the Late Proterozoic basins. These zones were favorable for generation of variable and abundant new evolutionary forms, some of which had principal advantages in competition for important biological resources of environments. As one can see, our contribution to this concept is quite concrete.

The last phase of reduction corresponding to the further development of regression was terminal in the generative stage and in the Miroedikha paleosuccession as a whole. At this phase (ninth and tenth levels), taxonomic composition of the "optimal" grouping was quickly reduced, and its distribution area degraded and disintegrated to local habitats located in sheltered shelf sectors. It is remarkable that in the course of the lateral expansion and the subsequent fragmentation of the distribution area during the regression (phases of mobilization and reduction), the "optimal" grouping experienced the influence of two evolutionary factors at least. In terms of neo-Darwinism (Timofeev-Resovsky *et al.*, 1977), these factors are the population waves and isolation, and it is well known that considerable changes in the abundance and taxonomic composition of recent biotic communities, as well as certain modifications of the genetic structure of populations, are related to effects of both factors (Mayer, 1968). Neontologic data like these stress once again a possible dependence of evolutionary modifications in the Miroedikha and other Riphean microbiotas on fluctuations of distribution areas of the "optimal" grouping (Veis *et al.*, 1998a, 1998b).

There was one more essential function of the reduction phase terminating the generative stage of the Miroedikha biota "ontogeny." This function prepared the "optimal" grouping to the crisis stage of biota evolution triggered by the predominance of shallow-water settings in the peritidal zones. The preparation included two concurrent phenomena: the appearance of refugia at the site of the regressing basin and the evacuation of a major mass of taxa of "optimal" and other, but shallow-water, groupings of microbiota to a vast buffer system of relevant zones of interrelated Precambrian marine basins. This concealed (infernal) phase or stage left no marks in paleontological record and is excluded so far from our model; however, it had to exist in order to secure the successive development of Proterozoic biotas.

Reviewers V.A. Krasilov and A.Yu. Rozanov

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Stratigraphy and Geological Correlation, Vol. 7, No. 1, 1999, pp. 35–48. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 7, No. 1, 1999, pp. 41–55.
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Late Paleozoic Radiolarians from the Polyfacies Formations of the Uralian Foredeep

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Received April 23, 1996; in final form July 10, 1997

Abstract—Stratigraphic and facies distribution of radiolarians within the interval from the Gzhelian Stage of the Upper Carboniferous to the Kungurian Stage of the Lower Permian in the Cis-Ural foredeep are considered. The interrelation between facies and specific features of radiolarian fauna is analyzed. In the longitudinal direction, the area of radiolarian distribution is limited by the foredeep boundaries. Its latitudinal interval coincides with the southern part of the foredeep below 57° N in the north. It is supposed that this latitude bordered the northern thermocline, which served as a paleoclimatic barrier to the dispersal of radiolarian biota northward.

Key words: *Upper Carboniferous, Lower Permian, the Urals, foredeep, facies, biostratigraphy, correlation, radiolarians, paleoecology.*

GENERAL CHARACTERISTICS OF THE CARBONATE–TERRIGENOUS DEPOSITS OF THE URALIAN FOREDEEP

The distribution of radiolarians in the Upper Paleozoic (Middle Carboniferous to Lower Permian) deposits is confined to the Uralian foredeep (UF), an enormous tectonic structure extended from the Caspian depression to Novaya Zemlya (Fig. 1). The UF was formed almost simultaneously in this vast territory during the Bashkirian Age and gradually moved to the west as the Paleourals grew. The greater part of the UF developed on the carbonate platform of the Visean–Serpukhovian age. Since the Bashkirian Age, a thick terrigenous wedge thinning out in the western direction was formed owing to the denudation of the thrust-and-fold rock complex. The wedge is underlain by coeval carbonate deposits wedging out to the east. The relationships between the carbonate and terrigenous deposits are schematically shown in Fig. 2.

There was no uniformity in the foredeep development. Some extensive areas, such as the Sylva and Yuryuzan'–Ai depressions (Fig. 1), migrated rapidly to the west. In these areas, not only the axis but also the entire foredeep were displaced, because the eastern part of the structure was involved into the thrust-and-fold dislocations and immediately denuded, as it is evident from the age of the carbonate boulder–pebble material.

Boundaries of other UF areas (for example, the Belaya and Aktyubinsk depressions) remained stable for a long time, and this resulted in the formation of thick (up to 1200 m) organogenic structures at the western edge of the UF and in the accumulation of a thick radiolarian-bearing clayey–siliceous member of the preflysch formation in the axial area.

The following facies succession, persistent from the east to the west (Figs. 3 and 5), was characteristic of the Middle Carboniferous–Early Permian time (Chuvashov, 1993, 1995):

(1) Predominantly sandy deposits, which always include the rudaceous sediments (gritstones, boulder–pebble conglomerates); this facies zone incorporates various organogenic buildups (hereinafter OB) with abundant and diverse assemblage of smaller foraminifers and fusulinids, corals, bryozoans, brachiopods, crinoids, calcareous algae, and stromatolites. Sandy turbidite flows originated at the margin of this shallow-water zone and moved deep into the basin. The term *coarse flysch* may be attributed to these deposits, which frequently include olistrostrome bodies. The existence time and width of this facies zone control the abundance of fossils in the next facies zone to the west.

(2) A relatively wide band of the *fine flysch* more persistent than the preceding zone is composed of regularly alternating sandstone, siltstone, and mudstone beds, which vary in thickness from 2 or 3 cm to 1 m. Some parts of the section are usually dominated by either clayey or sandy deposits. Calcareous–marlaceous and argillaceous–siliceous members also occur.

These deposits yielded two (allochthonous and autochthonous) fossil assemblages. The first assemblage is confined to the basal part of the thickest sandy or gravelly interlayers showing distinct graded bedding. Organic remains are concentrated within one fourth of the basal bedded interval of the interlayers,

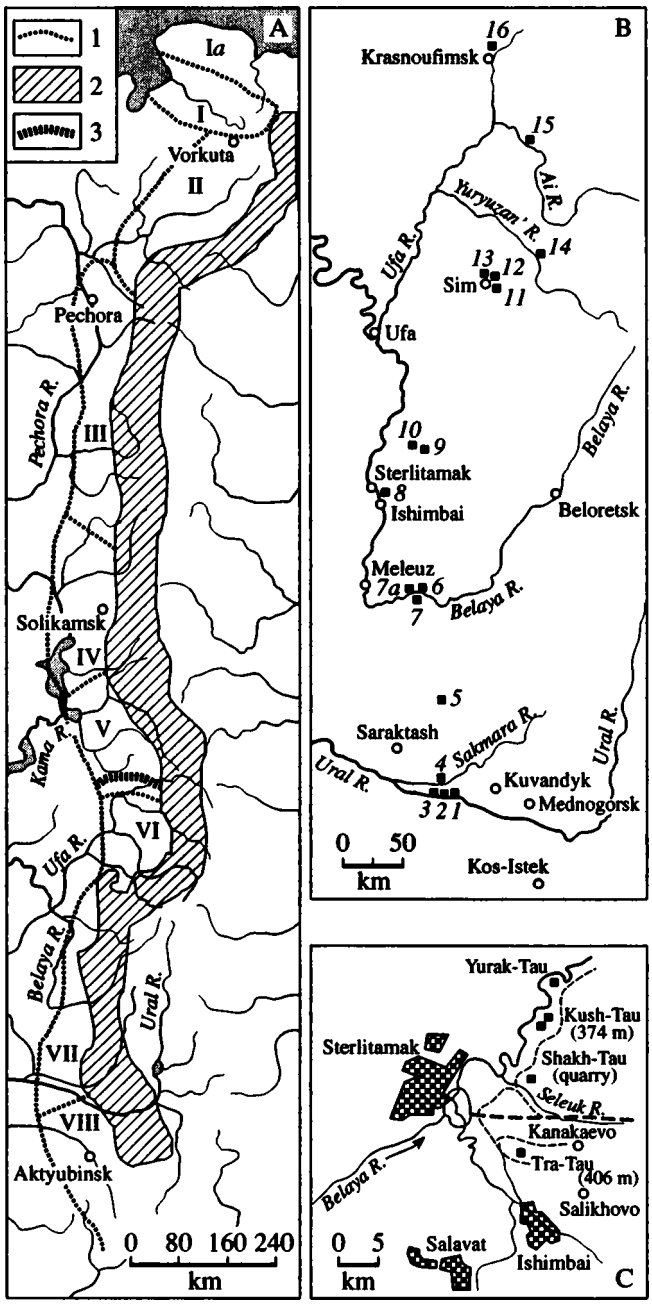

Fig. 1. Paleotectonic structures and sections studied.

A—The modern position of the Uralian foredeep, its boundaries and parts (depressions)

(1) western boundary of the foredeep; (2) deposits older than Permian; (3) northern limit of radiolarian distribution. Depressions: (I) Karataikha, (Ia) Kara, (II) Kos'va-Rogovaya, (III) Upper Pechora, (IV) Solikamsk, (V) Sylva, (VI) Yuryuzan'-Ai, (VII) Belaya–Sim, (VIII) Aktyubinsk.

B—Sections analyzed in this work: (*1*) Donskoe; (*2*) Nikol'skii; (*3*) Verkhneozernyi; (*4*) Kondurovka; (*5*) Shafeevka; (*6*) Siryat'; (*7*) Nizhnee Bikkuzino; (*7a*) Ishtuganovo; (*8*) Sterlitamak shikhany; (*9*) Usol'ka; (*10*) Tyul'kas; (*11*) near the town of Sim; (*12*) at the Sim railway station; (*13*) Biyanka; (*14*) Yuryuzan'; (*15*) Alegazovo; (*16*) Aleksandrovskoe.

C—Reef massifs near the town of Sterlitamak.

usually inside thin (1 or 2 cm) laminae. The most abundant fossils are smaller foraminifers and fusulinids, bryozoans, crinoids, and calcareous algae, whereas ammonoids and nautiloids, fragments of solitary and colonial corals and brachiopod shells are less frequent.

Mudstones, marls, and pelitomorphic limestones separating the turbidites contain remains of other groups of organisms: radiolarians, sponge spicules, small solitary corals, thin-shelled brachiopods (some inarticulate), and pelecypods. Frequently, these sediments are completely reworked by mud-eaters. Ammonoids and orthoceratids likely inhabited the clayey and marlaceous sediments, where all evolutionary stages of these organisms, from embryonic to adult species, are recorded.

(3) To the west, the fine flysch is replaced by a thin member of siliceous mudstone, chert, mudstone, and marl. The member includes interbeds of micrite, grainstone, and carbonate breccia. The fossil assemblage of this facies zone is very similar to that from the thin flysch members between turbidites. Spicule-bearing interbeds occur at the base of this member known under the names *preflysch formation* or *preflysch*. Radiolarians occur higher in the section, near the flysch base.

(4) The preflysch formation may immediately contact the zone of organogenic buildups or be separated from it by a relatively narrow zone of bedded limestone, dolomite, and carbonate breccia. The total thickness of beds in the latter is several times less than that of the organogenic buildups and approximates the thickness of the preflysch formation. Sediments of this type may be referred to the *prereef facies*. They are poorly studied in few boreholes with insufficient core recovery. Radiolarians were found in the prereef marls and pelitomorphic limestones.

(5) Next is the *reef facies*. Up to 12 belts of organogenic buildups different in age, which marked the geomorphologic and facial platform–foredeep boundary in the Late Paleozoic (Fig. 3), can be distinguished in this case. The thickness and ecological maturity of the OB are controlled by the stability of this boundary. Radiolarians usually do not occur in the reef and western facies. An exception is the case when clayey or carbonate-clayey beds are intercalated with reef deposits in response to a certain displacement of the platform–foredeep boundary, and then radiolarians occur in the frontal parts of the reef massifs.

Distribution patterns of radiolarian facies at the western edge of the UF may be supplemented with data on the Karachaganak oil field in the northern Caspian depression (Afanas'eva *et al.*, 1986; Afanasieva and Zamilatskaya, 1993). The Karachaganak reef massif is an atoll with an inner lagoon, where radiolarian-bearing fine-grained limestones were sporadically accumulated. Radiolarians also occur in thin beds of marlaceous–clayey sediments of the Karachaganak prereef zone and in the basinal facies remote from the massif.

A brief review shows that distribution areas of clay, carbonate, and siliceous fine-grained muds were most rich in radiolarians. Radiolarians occur in all sedimentary facies, including those of the shallow-water zones of the coarse flysch (in the east), prereef sediments, and lagoonal deposits.

There is a certain trend in the latitudinal distribution of radiolarians. The northernmost locality of radiolarians is in the Ufa River basin (Fig. 2). To the north of it, neither we, nor other geologists have found radiolarians. As is evident from regular changes in the compositions of all fossil groups of the Early Permian basin from south to the north (Chuvashov, 1991, 1995), the confinement of radiolarians to the southern UF part is supposed to be the result of a climatic factor, primarily, of a higher water temperature than in the northern part.

STRATIGRAPHIC AND FACIES DISTRIBUTION OF RADIOLARIANS

By the present time, radiolarians have been found in the large (over 1000 km) area of the southern UF part and in the south of its middle part (Fig. 1).

There is some information on occurrences of radiolarians in the Middle–Upper Carboniferous and Lower Permian carbonate–terrigenous deposits of the Bol'shoi Ik River basin exposed to the north of the Sakmara River (Khvorova, 1961). However, their taxonomic composition has not been studied yet.

Nazarov revealed the following stratigraphic succession of beds with radiolarians in the interfluve between the Ural and Sakmara rivers, i.e., in the southernmost area of the territory under consideration (Isakova and Nazarov, 1986; Nazarov, 1988; Nazarov and Ormiston, 1985).

(1) The *Tormentum pervagatum* assemblage was identified in beds 14–27 of the Nikol'skii section. This interval corresponds to the lower part of the *Daixina sokensis* Zone of the Gzhelian Stage (Ruzhentsev, 1950; Davydov and Popov, 1991).

Radiolarians from the Carboniferous–Permian boundary deposits were studied in the Nikol'skii section only, where the Lower Permian boundary is recorded at the base of Bed 41. This level is quite close to the boundary between the Carboniferous and Permian systems in the Unified Stratigraphic Scheme for the Permian of the Urals (Chuvashov *et al.*, 1994).

(2) The *Tormentum protei* assemblage is characteristic of the upper Gzhelian Stage corresponding to beds 28–40 of the Nikol'skii section. In general, this interval corresponds to the upper part of the fusulinid *Daixina sokensis* Zone, whereas the uppermost Bed 40 is correlative to the *Daixina bosbytauensis* Zone (Davydov and Popov, 1991).

(3) The *Latentifistula crux* assemblage was found in beds 41–44 of the same section, which correspond to the lower part of the *Daixina bosbytauensis–Daixina robusta* Zone.

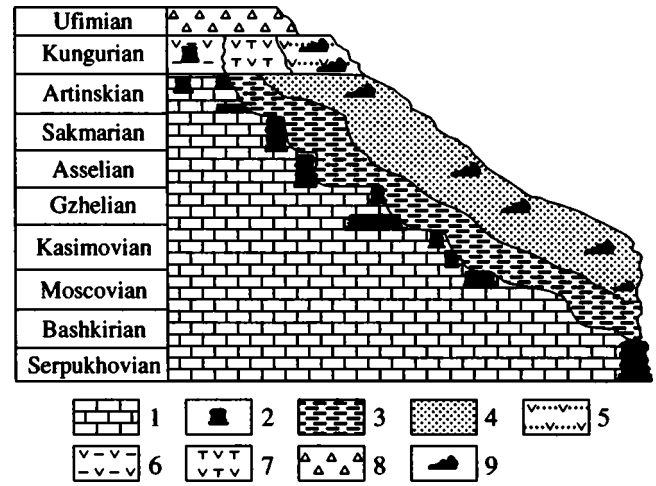

Fig. 2. Age and structural relationships between carbonate and terrigenous formations of the Uralian foredeep: (1) carbonate formation, subformation of bedded carbonates; (2) carbonate formation, subformation of organogenic buildups; (3) preflysch formation; (4) flysch formation; (5) evaporite formation, terrigenous shallow-water subformation; (6) evaporite formation, deep-water subformation; (7) evaporite formation, shallow-water carbonate–evaporite subformation; (8) terrestrial red molasse formation; (9) nearshore organogenic buildups.

The Asselian radiolarians are inadequately studied. Some preliminary information on the issue was published by Isakova and Nazarov (1986).

(4) The Tormentidae assemblage is arbitrarily assigned to the *Pseudoschwagerina moelleri–Pseudofusulina fecunda* Zone.

(5) The *Copycintra* sp. assemblage is supposed to characterize the Shikhan Horizon of the Asselian Stage and the lower Tastuba Horizon of the Sakmarian Stage.

The representative radiolarian fauna, which we named the *Tetragregnon vimineum–Copiellintra diploacantha* assemblage, was found in the *Sphaeroschwagerina sphaerica–P. firma* Zone of some sections near the town of Orenburg. The stratotype of the *Tetragregnon vimineum–Copiellintra diploacantha* Beds is the Kondurovka section at the right bank of the Sakmara River near the "Kondurovka" railway station. This section is also the stratotype of the Sakmarian Stage (Chuvashov *et al.*, 1991). The mentioned radiolarian assemblage was recovered from the samples taken 2 m above the base of Bed 1 of Exposure 2015 in the southern extremity of the Nos Mountain, the Karamuruntau Ridge.

The *Tetragregnon vimineum–Copiellintra diploacantha* assemblage consists of the following species: *Entactinia* sp., *E.* cf. *pycnoclada* Naz. et Orm., *Entactinosphaera* sp., *E. crassicalthrata* Naz. et Orm., *E. calthrata* Naz. et Orm., *Astroentactinia luxuria* Naz. et Orm., *Tetragregnon vimineum* Amon et Braun, *Helioentactinia biexosphaera* Naz. et Orm., *Copucyntra* sp., *C. cuspidata* Naz. et Orm., *C. phymatodonta*

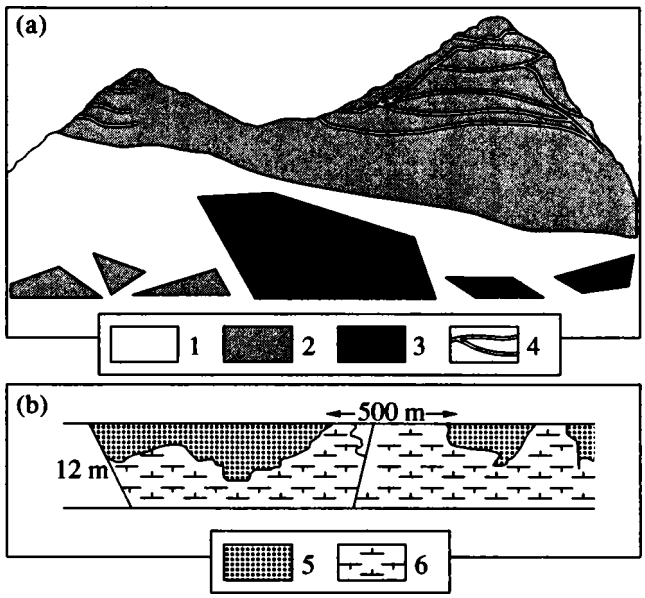

Fig. 3. Radiolarian-bearing micrite interbeds within the carbonate Yurak-Tau reef massif (a) and the contact between the Artinskian clayey–marlaceous deposits and Sakmarian reef carbonates in the Shakh-Tau section (b). Yurak-Tau section: (1) unexposed talus areas; (2) reef limestone; (3) bedded limestone; (4) beds with radiolarians; Shakh-Tau section: (5) Artinskian clayey–marlaceous deposits; (6) Sakmarian reef carbonates.

Naz. et Orm., *Copiellintra* sp., *C. diploacantha* Naz. et Orm., *Polyentactinia* sp. (sp. nov.), *Tormentum* sp., and *T. circumfusum* Naz. et Orm. The characteristic feature of the assemblage is the predominance (up to 80% of the assemblage) of large spherical form *Helioentactinia biexosphaera* Naz. et Orm. On the basis of some indirect evidence, the *Copicyntra* sp. assemblage was suggested to be the facial analogue of the *Tetragregnon vimineum–Copiellintra diploacantha* assemblage. The former is characteristic of shallow-water deposits, whereas the latter mainly occurs in deeper facies. The taxonomic composition of the first assemblage is less diverse as compared to the second one. In addition to radiolarians, the same samples yielded conodonts attributed by Chernykh to the *Mesogondolella pseudodostriata* Zone. Their assemblage is characteristic of some sections in the Cis-Ural region near the town of Orenburg.

(6) The radiolarian assemblage from the upper Sarabil' Formation of the Sakmarian stratotype (the *Pseudofusulina verneuili* fusulinid zone after Chuvashov *et al.*, 1991) was studied by Kozur (1981). Two new genera were identified in this assemblage. The list of forms from the Sarabil' Formation stratotype is as follows: *Parafollicuculus sakmarensis* Kozur, *P.(?) nazarovi* Kozur, *Spinodeflandrella tetraspinosa* Kozur, *Holdsworthella permica* Kozur, and *H. perforata* Kozur. This list can be supplemented with two new species of genus *Nazarovispongus* (*N. pavlovi* Kozur and *N. permicus* Kozur), both described by Kozur (1980) from the Sarabil' Formation section near the Verkhneozernaya Settlement, and with *Entactinia pycnoclada* Naz. et Orm., *Helioentactinia* sp., and *Haplodiacanthus* cf. *rombothoracata* (Ishiga et Imoto), the forms identified by Nazarov and Ormiston (1985) in the same section.

(7) Next radiolarian-bearing level is recorded in the Malyi Ik Formation of the Sterlitamak Horizon of the Sakmarian Stage, where Nazarov (Isakova and Nazarov, 1986) reported the occurrence of *Latentifistula triacantophora* Naz. et Orm. and *Tormentum circumfusum* Naz. et Orm. The Kondurovka Formation of the upper Sterlitamak Horizon contains *Ruzhencevispongus* (?) *plumatus* Naz. et Orm., *Copiellinta diplocantha* Naz. et Orm., and *Latentifustula valdeinepta* Naz. et Orm. (Nazarov and Ormiston, 1985). The assemblages from the Malyi Ik and Kondurovka formations correspond to the level of fusulinid assemblage of the *Pseudofusulina urdalensis* Zone.

(8) The diverse radiolarians of the *Entactinosphaera crassiclathrata–Quinqueremis arundinea* assemblage occur in the Aktasty Formation of the Artinskian Stage. This formation is correlated to the Burtsevka and Irgina horizons of the stage. After investigation of the Aktasty Formation section near the Donskoe Settlement, Nazarov and Ormiston (Nazarov and Ormiston, 1985; Nazarov, 1988) listed the following radiolarians characteristic of the *Entactinosphaera crassiclathrata–Quinqueremis arundinea* Zone: *Entactinosphaera crassiclathrata* Naz. et Orm., *E. strangulata* Naz. et Orm., *Ruzhencevispongus cataphractus* Naz. et Orm., *R. retiporosus* Naz. et Orm., *R. partilaminatus* Naz. et Orm., *R.(?) eacumenireticulatus* Naz. et Orm., *R.(?) eamarginatus* Naz. et Orm., *R.(?) promiscus* Naz. et Orm., *Latentifistula neotenica* Naz. et Orm., *Quinqueremis arundinea* Naz. et Orm., *Quadriremis gliptoacus* Naz. et Orm., *Tetratormentum narthecium* Naz. et Orm., *Rectotormentum fornicatum* Naz. et Orm., *Albaillella apporecta* Naz. et Orm., *Haplodiacanthus* sp., *Astroentactinia luxuria* Naz. et Orm., *Polyentactinia fiscina* Naz. et Orm., *P. fragilis* Naz. et Orm., *P. affirmata* Naz. et Orm., and *P. applanata* Naz. et Orm.

At the Aktasty River, the section of the Baigendzhinian Substage (the Sarga and Sarana horizons) of the Artinskian Stage yielded *Polyentactinia lautitia* Naz. et Orm., *Copicyntra phymatodonta* Naz. et Orm., and *C. cuspidata* Naz. et Orm.

Radiolarians from the Shafeevka Formation of the upper Sarga Horizon of the Artinskian Stage are of interest in the paleoecological aspect. At the Shafeevka River in the Syuren' River basin, the following section is exposed (Chuvashov *et al.*, 1990):

(1) The lower Sarga Horizon of the Artinskian Stage is represented by a thick sequence of Baigendzhinian boulder–pebble conglomerates with lenses and members of gritstones and gravelly sandstones.

(2) The overlying Shafeevka Formation is represented by *Tubiphites*–stromatolitic bioherms, which

locally include lenticular interbeds of stromatolites. The reddish-brown marl beds occur between the bioherms. Bodies of the Shafeevka marls are highly fossiliferous. Interbeds of sponge spicules, brachiopods, ammonoids, nautiloids, conodonts, and less frequent radiolarians of the *Polyentactinia lautitia* Zone (two latter fossil groups occur only in the upper Artinskian part of the formation) are recorded at many levels. Radiolarian shells are insufficiently preserved, frequently recrystallized, and replaced by calcite. The assemblage is dominated by large spherical forms. The mentioned biohermal buildups were formed in close proximity to the shore, and, hence, this oryctoassemblage of radiolarians can be referred to the littoral shallow-water type.

Some sections exhibiting numerous radiolarian levels were studied in the Belaya River basin. In general, the Carboniferous–Permian section at the right bank of the river between the Kuznetsovskii and Syrtlanovo villages (the Siryat section in Chuvashov *et al.*, 1990) exhibits the following succession of rock units:

(1) The Bashkirian Stage: shallow-water limestones and dolomites enclose numerous chert concretions bearing the abundant and diverse fauna of fusulinids, rugose corals, tabulates, brachiopods, and crinoids. Calcareous algae occur as well. No radiolarians were found. The upper Bashkirian Substage is 76 m thick.

(2) The Moscovian Stage, the lower Moscovian Substage. The lower member is represented by thick- to thin-bedded fine-grained and pelitomorphic limestones, which are irregularly dolomitized and include numerous lenticular interbeds and concretions of cherts; this member contains grainstone interbeds with fusulinids, bryozoans, and crinoids. Corals, brachiopods, and crinoids occur throughout the interval; no radiolarians were found. The upper member of the substage (15 m thick) is composed of steel-gray bedded pelitomorphic limestones with numerous lenticular chert interbeds; organic remains are represented here by rare foraminifers and abundant crinoid columnals at some levels. The distinct impoverishment of the fauna upsection can be explained by the gradual deepening of the basin. The lower Moscovian Substage is 38 m thick.

(3) The upper Moscovian Substage of the Upper Carboniferous and the Asselian Stage or the Tastuba Horizon of the Sakmarian Stage. This thin member corresponding to a large stratigraphic interval is attributed to the preflysch formation. It was accumulated at the initial period of radiolarian dispersal in the basin. The preflysch formation is composed of dark-gray mudstones, flaggy marls, pelitomorphic limestones, and dolomites; cherts commonly occur in the lower half of the member. A large part of the section contains interbeds and concretions of phosphorite. Organic remains are represented by radiolarians, small thin-shelled brachiopods, gastropods, ammonoids, scales and bones of fishes, and worm tracks. Radiolarians were found in pelitomorphic limestones and dolomites, as well as in the phosphorite concretions. The thickness is 28 m.

(4) The Sterlitamak Horizon of the Sakmarian Stage. The unit is represented mainly by sandstones with interbeds of siltstones and mudstones. There are grainstone layers up to 0.7 m thick showing the distinct graded bedding. These deposits include subordinate members of alternating mudstones and sandstones, in which clay sediments considerably prevail. The mudstone beds contain interbeds of pelitomorphic limestones and marls, many of which bear shells of radiolarians of the *Entactinia pycnoclada–Tormentum circumfusum* Zone. The unit thickness is 151 m.

(5) The Burtsevka and Irgina horizons, and the lower part of the Sarga Horizon. These deposits correlative to the Aktasty Formation are represented by alternating sandstone and mudstone beds. The lower parts of the sandstone beds with distinct graded bedding frequently contain abundant fusulinids, bryozoans, crinoids, and algae, which were redeposited from the shallow-water zone. Some intervals of the section are dominated by clay deposits with numerous interbeds of pelitomorphic limestones and marl members. Rare radiolarians coexisting with ammonoids and pelecypods are confined, as a rule, to limestone and marl beds. Radiolarians represent a succession of *Rectotormentum fornicatum, Entactinosphaera crassiclathrata–Quinqueremis arundinea*, and *Teracircinata reconda* assemblages. The unit is 1153 m thick.

(6) The upper Sarga Horizon is represented mainly by massive thick-bedded and, sometimes, gravelly sandstones with admixture of small pebbles. The sandstone beds include interbeds of mudstone and siltstone, which locally contain large concretions of clayey limestone with ammonoids. No radiolarians were found in this interval. The apparent thickness is 90 m.

The lower part of this succession can be observed in natural exposures and a quarry of the Mt. Bala-Tau, the left bank of the Belaya River, 2 km to the south of the right-bank section described. In the Mt. Bala-Tau section, the unexposed interval ten meters wide is succeeded by the upper Bashkirian limestones, which are overlain by the following units:

(1) Gray and dark-gray bedded (5–15 m) pelitomorphic, locally dolomitized limestones with numerous chert concretions; the sediments are heavily reworked by worms. No radiolarians were found. The thickness is 8.7 m.

(2) Alternating beds of gray fine-grained dolomites, greenish gray mudstones, and dolomitized marls; the unit also includes lenticular chert interbeds. All sediments are heavily bioturbated. Abundant recrystallized radiolarian shells occur in the marl and chert beds, but their taxonomic composition has not been studied. The thickness is 3 m.

(3) The member incorporating nine tuff interbeds ranging in thickness from 1 or 2 to 15 cm; in two samples of the rocks (L22BA and L25BA), M. Caridroit

identified radiolarians *Pseudoalbaillella elegans* Ishiga et Imoto; *Pseudoalbaillella* sp. 1 Caridroit, 1986; *Nazarovella gracilis* De Wever et Caridroit; *Latentiffistula* aff. *triacantophora* Nazarov et Ormiston; *Tormentum* aff. *circumfusum* Nazarov et Ormiston; *Latentibifistula* sp., *Quinqueremis* sp.; and *Entactinia* sp. The thickness is 4.2 m.

The assemblage of listed radiolarians allows us to assign Beds 2 and 3 to the Gzhelian–upper Asselian. This is in accordance with the occurrence of the late Asselian fusulinids in carbonate–terrigenous deposits 42 m higher in the section.

Thus, an intense dispersal of radiolarians over biotopes of the preflysch and flysch formations of the Belaya River basin (the Siryat' and Bala-Tau sections) began in the late Moscovian time and terminated in the Artinskian Age. Abundant data on other fossil groups, along with the results of the radiolarian analysis, permit the conclusion that the radiolarian dispersal was stimulated by the gradual deepening of the basin and the influx of the Tethyan warm-water masses into the territory under consideration. Against this background, the bursts of radiolarian development can be related to the periods of volcanic activity (the tuff-bearing member of the Bala-Tau section).

Distribution patterns of radiolarians in the area of large reef massifs near the town of Sterlitamak is of interest for understanding trends in their dispersal (Figs. 2 and 3). The development of reefs ceased in the mid-Asselian time in the eastern part of the area (the Tra-Tau Mountain) and in the early Artinskian time in the western one. Owing to erosion, a complicated relief with an amplitude gradient of few hundred meters was formed. Afterwards, the reef massifs rapidly subsided, and their uneven surface was buried under marlaceous–clayey sediments of variable thickness. The type of sedimentation was dependent on the basin depth. Red shallow-water siliceous marls with abundant sponge spicules overlie the Asselian reef deposits of the Tra-Tau and Sterlitamak areas, and the lower Artinskian reef limestones of the Yurak-Tau site. Meanwhile, the reef massif under consideration demonstrates numerous interbeds of gray radiolarian micrite (Fig. 3) reflecting episodes of the short-time abrupt subsidence of the massif, pauses in the regular reef-forming process, and influx phases of pelagic waters with the typical marine plankton. The taxonomic composition of radiolarians from these interbeds has not been studied yet.

In the upper escarpment of the Shakh-Tau massif (Fig. 3) exposed in a quarry, the uneven surface of the Sterlitamak reef limestones is covered with dark-gray and brownish gray mudstones and marls with thin manganese interbeds. Shells of radiolarians, which have not been studied in the taxonomic aspect, are abundant in some marl interbeds. This data indicates that radiolarians appear only at a certain depth of a deepening basin, and often the first inhabitants are sponges. Radiolarians are frequently associated with ammonites.

Radiolarians characterize almost entire Lower Permian section of inconsiderable thickness near the Krasnousol'skii Settlement (Fig. 4). The section comprises two outcrops supplementing each other (at the Usol'ka and Tyul'kas rivers; Chuvashov *et al.*, 1983, 1990) and exposures of quarries and road excavations. The composite section can be briefly described in the following way (from the base upward):

(1) At the Usol'ka and Tyul'kas rivers, the basal part of the clayey–siliceous–marlaceous sequence (the preflysch formation) includes limestone and dolomite beds of the upper Moscovian Subage. This section is located to the west of the above-described succession at the Belaya River in the area between the Kuznetsovskii and Siryat' villages.

(2) The Myachkovo Horizon of the Moscovian Stage–Kasimovian Stage of the Upper Carboniferous. The lower part of the unit is represented predominantly by dark-gray and black mudstones with interbeds of siliceous mudstones and pelitomorphic highly silicified limestones. The upper part shows the prevalence of pelitomorphic silicified limestones and dark-gray dolomites over the intercalated mudstones and dolomitic marls. Organic remains consist of smaller foraminifers and fusulinids (in thin detritic lenses inside thicker limestone interbeds), radiolarians (in cherts and marls), sponge spiculas (in silicified limestones), solitary corals, and small brachiopods. Thickness is 13 m.

(3) The Gzhelian Stage of the Upper Carboniferous. The most part of the stage is represented by dark-gray thin-bedded mudstones with thin interbeds of brownish fine-grained dolomite and dolomitic marl, and pelitomorphic limestone. Small (2–10 cm) phosphorite and large (up to 25 cm long) carbonate concretions are common in mudstones. In the middle part of the stage, there is a horizon of slided carbonate blocks containing abundant fusulinids, corals, and brachiopods. At many levels, dolomites and dolomitic marls bear radiolarians, sponge spicules, small brachiopods (including their inarticulate species), pelecypods, nautiloids, trilobites, conodonts, and fish scales. The thickness is 15 m.

(4) The Asselian Stage of the Lower Permian, the Kholodnyi Log Horizon. The lower part of the horizon is predominantly composed of brownish gray flaggy mudstones, which include, with an interval of 0.5–1 m, interbeds of dolomite or pelitomorphic limestone. Phosphorite concretions are widespread in mudstones. The upper part of the horizon shows the reverse proportion of the rock abundance: the dark-gray beds of pelitomorphic limestone dominate and the mudstone interbeds are reduced in thickness. Organic remains are represented by radiolarians (in dolomites, marls, and phosphorites), sponge spicules (in limestones), corals, inarticulate and small articulate brachiopods, ammonoids, nautiloids, small bones and scales of fishes, and conodonts. The horizon is 22 m thick.

In two closely spaced samples from the upper part of the horizon, Caridroit identified radiolarians

Pseudoalbaillella bulbosa Ishiga, *P. nodosa* Ishiga, *P. U-forma* Holdsworth et Johns (1980), and other representatives of Albaillellidae, Latentifistulidae, and Entactinidae families.

(5) The Asselian Stage, Shikhany Horizon. The horizon is predominantly composed of seel-gray pelitomorphic limestones. Smaller foraminifers, fusulinids, detritus of bryozoans, and brachiopods, together with algae, were found in basal parts of the thickest beds. Fine-grained carbonate breccias occur as well in the units. The limestones are bedded. The horizon is 14 m thick.

(6) The Sakmarian Stage, Tastuba Horizon. The horizon is mainly composed of brownish-gray flaggy marls frequently microbedded; it also includes interbeds of steel-gray pelitomorphic limestone. The upper part of the horizon shows graded bedding in limestones, whose basal laminae contain fusulinids, bryozoans, and crinoids. Marls bear abundant recrystallized (calcitized) radiolarian shells. The horizon is 10 m thick.

(7) The Sakmarian Stage, Sterlitamak Horizon. The horizon is similarly constituted by brownish gray, flaggy, frequently microbedded marls irregularly intercalated with interbeds of steel-gray pelitomorphic limestone (micrite). Many marl interbeds can be taken for radiolarites. At some levels of the marls, there are pelecypod coquinas. Abundant radiolarian shells are usually highly recrystallized and calcitized. The incomplete thickness (an effect of dislocation) is 38 m.

The Artinskian deposits are described as they are exposed at the Tyul'kas River.

(8) The Artinskian Stage, Burtsevka Horizon. At the base of the horizon, there is a carbonate breccia whose lower part is composed of angular clasts, and, less frequently, of boulders and large pebbles of pelitomorphic and detrital limestones. Higher, the breccia grades into pelitomorphic flaggy limestone. The breccia cement contains abundant fusulinids, whereas the pelitomorphic limestone yielded conodonts. The thickness is 1.5 m.

The major part of the unit is represented by greenish and yellowish gray flaggy marls with rare interbeds of mudstones and clayey siltstones of the same color. In this rather thick sequence, there are lenses, lenticular interbeds, and beds of pelitomorphic limestone 5–15 cm thick and spaced for 0.8–2 m one from another. The thickest beds show the basal detrital part (1–5 cm thick) containing fusulinids, bryozoans, and crinoids. Many interbeds of pelitomorphic limestone contain abundant radiolarians of the *Entactinosphaera crassiclathrata–Quinqueremis arundinea* Zone (identifications of Amon): *Entactinia densissima* Naz. et Orm., *Entactinosphaera crassiclathrata* Naz. et Orm., *E. strangulata* Naz. et Orm., *Astroentactinia luxuria* Naz. et Orm., *Helioentactinia biexosphaera* Naz., *Copicyntra cuspidata* Naz. et Orm., *C. phymatodonta* Naz. et Orm., *Copiellintra* sp., *Tetracircinata reconda* Naz. et Orm.,

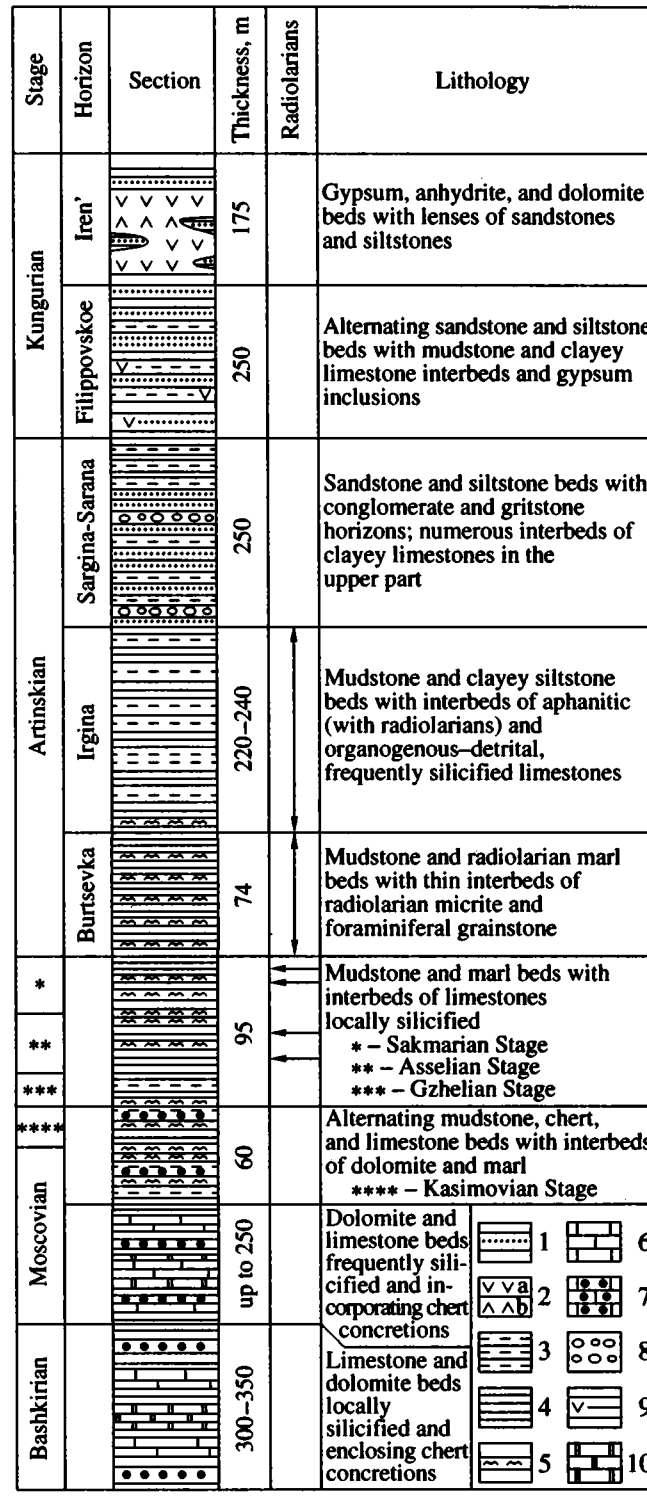

Fig. 4. The Upper Paleozoic section in the Krasnousol'sk region: (1) sandstone; (2a) gypsum; (2b) anhydrite; (3) siltstone; (4) mudstone; (5) marl and clayey limestone; (6) limestone; (7) chert concretions; (8) conglomerate; (9) gypsinate deposits; (10) dolomite (arrows point to levels with radiolarians; out of scale).

Ruzhencevispongus cataphractus Naz. et Orm., *R. retiporosus* Naz. et Orm., *Latentifistula neotenica* Naz. et Orm., *Quinqueremis arundinea* Naz. et Orm., *Quadriremis gliptoacus* Naz. et Orm., *Tetratormentum nartheecium* Naz. et Orm., *Rectotormentum fornicatum* Naz. et Orm., *Albaillella apporecta* Naz. et Orm., *Haplodiacanthus* sp., *H. anfractus* Naz. et Rud., *Polyentactinia fiscina* Naz. et Orm., and *P. aranaeosa* Naz. et Orm. The horizon is about 75 m thick.

(9) The Artinskian Stage, Irginian Horizon. The major (background) part of the section is represented by greenish-gray mudstones, silty mudstones, and marls with thin interbeds of pelitomorphic limestone. There are radiolarians (in marls and limestones) and thin-shelled small brachiopods and pelecypods (in mudstones and marls). These deposits include interbeds of fusulinid–crinoid–bryozoan grainstones, which are 20 cm thick and spaced by intervals of 1–10 m in the lower part of the horizon; similar interbeds are up to 1 m thick and more frequent in the upper part. The horizon is 130 m thick.

(10) No radiolarians were found in the overlying predominantly sandy formations with interbeds of gritstones and conglomerates, which are referred to the Sargina and Sarana horizons.

In the Sim area, several Middle Carboniferous–Lower Permian sections were studied; they show the following bed succession (from the base upward, according to Chuvashov *et al.*, 1990):

The Kolosleikino Formation spans the upper part of the Bashkirian Stage and entire Moscovian Stage of the Upper Carboniferous, and the lower part of the Kholodnyi Log Horizon of the Asselian Stage. The formation lies on the middle limestone beds of the lower Bashkirian Substage. It is composed of dark-gray siliceous mudstones and cherts with interbeds of marl and gray pelitomorphic and, less frequently, detrital, limestones. Organic remains include radiolarians, small thin-valved brachiopods, pelecypods, ammonoids, gastropods, scales and bones of fishes, and conodonts. The thickness is up to 50 m.

This unit of inconsiderable thickness characterizes the condensed succession of beds spanning the broad stratigraphic interval, as is evident from the occurrences of Bashkirian conodonts at the base; the detrital limestone interbeds with Moscovian fusulinids in the middle part; and Late Carboniferous conodonts, ammonoids, and fusulinids in the upper part of the section (Chuvashov *et al.*, 1990). The unit belongs to the preflysch formation.

The Domennaya Gora Formation spans the upper part of the Asselian Stage (the fusulinid *Sphaeroschwagerina moelleri–Pseudofusulina fecunda* and *Sphaeroschwagerina sphaerica–Pseudofusulina firma* zones) and the lower part of the Tastuba Horizon of the Sakmarian Stage (the *Pseudofusulina moelleri* Zone). The formation is composed of thick interbedded mudstones, marls, and clayey pelitomorphic limestones, which include beds (0.4–3 m thick and spaced for 3–10 m) of graded fine-grained carbonate–siliceous breccias, gritstones, or sandstones. In the section to the south of the town of Sim, there are several horizons of breccias with carbonate blocks, whereas the section near the Sim railway station is dominated by calcareous boulder–pebble conglomerates. The gritstone and sandstone beds with graded bedding contain diverse shallow-water fauna of fusulinids, solitary and fragmented colonial corals, brachiopods, bryozoans, crinoids, and calcareous algae. Marls, mudstones, and micrites between these beds are characterized by radiolarians, sponge spicules, nautiloids, ammonoids, and fish remains. The formation thickness is up to 500 m. This unit terminates the preflysch formation.

The Pravoberezhnaya Formation corresponds to the upper part of the Tastuba Horizon (the fusulinid *Pseudousulina verneuili* Zone) and the Sterlitamak Horizon of the Sakmarian Stage. The formation is constituted mainly by the alternation of polymictic sandstones and gritstones, on the one hand, and mudstones and siltstones, on the other hand. Upsection, the sandstone interbeds increase in thickness (locally up to 0.7–1 m) and become more frequent. The sandstone interbeds show distinct graded bedding. Locally, the most coarse-grained lower parts of the interbeds bear fusulinids, crinoids, and bryozoans. Interbeds and members of marl contain radiolarians, ammonoids, and nautiloids, as well as brachiopods in some places. Radiolarians and cephalopods occur also in carbonate concretions and micritic interbeds. The Tastuba deposits contain radiolarians of the *Helioentactinia ikka–Haplodiacanthus perforatus* Zone, such as *Entactinia dolichoacus* Naz., *Entactinosphaera aenigma* Naz., *Astroentactinia insecta* Naz., *Helioentactinia ikka* Naz., *Copicyntra aciloxa* Naz., *Tetragregnon spharicus* Naz., *Latentifistula torulosa* Naz., *L. crux* Naz. et Orm., *Latentibifistula* sp., *Tormentum aequilateralis* Naz., *Tetratormentum crateris* Naz., *Albaillella permica* (Kozur), *Haplodiacanthus* sp., *H. perforatus* (Kozur), etc. (40 taxa in total). The Sterlitamak part of the formation contains the following radiolarians typical of the *Entactinia pycnoclada–Tormentum circumfusum* Zone: *Entactinia pycnoclada* Naz. et Orm., *Entactinosphaera strangulata* Naz. et Orm., *Helioentactinia biexosphaera* Naz., *Latentifistula valdeinepta* Naz. et Orm., *Latentibifistula triacanthophora* Naz. et Orm., *Polyfistula longiquitas* Naz. et Orm., *Ruzhencevispongus plumatus* Naz. et Orm., *R. uralicus* Kozur, *Tormentum circumfusum* Naz. et Orm., *Haplodiacanthus sakmaricus* (Kozur), *Camptoalatus* sp., etc. (36 taxa in total; Amon and Chuvashov, 1992). The flysch formation (the fine flysch) begins with the basal beds of this formation. The formation is up to 450 m thick.

The upper constituent of the section is the Shelyvaginskaya Formation consisting of the Burtsevka and Irgina horizons of the Artinskian Stage. The lower part of the formation (the Burtsevka Horizon) is composed of sandstones with marl interbeds, and its upper part

(the Irgina Horizon) is constituted by alternating conglomerate and sandstone members. The conglomerate beds of the section include thick (up to 40 m) subordinate members of thin-bedded flaggy marls bearing sponges. Radiolarians occur in abundance in thin interbeds and lenses of micrite of the Burtsevka Horizon, i.e., in that interval of the Shelyvaginskaya Formation, which corresponds to the fusulinid *Pseudofusulina concavutas–Pseudofusulina pedissequa* Zone (Amon and Chuvashov, 1992; Amon *et al.*, 1990). The radiolarian assemblage of the *Entactinosphaera crassiclathrata–Quinqueremis arundinea* Zone from this interval includes *Entactinia densissima* Naz. et Orm., *E. spinifera* Amon et Braun, *E. faceta* Amon et Braun, *Entactinosphaera crassiclathrata* Naz. et Orm., *E. strangulata* Naz. et Orm., *E. calthrata* Naz., *Astroentactinia luxuria* Naz. et Orm., *Spongentactinia simensis* Amon et Braun, *Spongentactinia rigida* Amon et Braun, *Helioentactinia biexosphaera* Naz., *Copicyntra cuspidata* Naz. et Orm., *C. phymatodonta* Naz. et Orm., *Copiellintra* sp., *Tetracircinata reconda* Naz. et Orm., *Ruzhencevispongus cataphractus* Naz. et Orm., *R. retiporosus* Naz. et Orm., *R. apertus* Amon et Braun, *Latentifistula neotenica* Naz. et Orm., *Quinqueremis arundinea* Naz. et Orm., *Quadriremis gliptoacus* Naz. et Orm., *Tetratormentum narthecium* Naz. et Orm., *Rectotormentum fornicatum* Naz. et Orm., *Albaiellella apporecta* Naz. et Orm., *Haplodiacanthus* sp., *H. anfractus* Naz. et Rud., *Polyentactinia fiscina* Naz. et Orm., *P. aranaeosa* Naz. et Orm., etc. (50 taxa in total).

A radiolarian assemblage of higher stratigraphic position was found in the northern Sim depression, 15 km to the north of the town of Sim. One of the boreholes drilled by the Southern Urals State Geological Enterprise recovered the well-preserved scarce radiolarians from the Irgina Horizon of the Shevalyginskaya Formation. Amon identified them as species of the *Tetracircinata reconda* Zone: *Entactinia pycnoclada* Naz. et Orm., *E. faceta* Amon et Braun, *Entactinosphaera calthrata* Naz., *Pluristratoentactinia* sp., *Astroentactinia luxuria* Naz. et Orm., *A. inscita* Naz., *Helioentactinia biexosphaera* Naz., *Spongentactinia* sp., *Copicyntra* sp., *C. phymatodonta* Naz. et Orm., *Ruzhencevispongus cataphractus* Naz. et Orm., and *Polyentactinia fiscina* Naz. et Orm. This assemblages is accompanied by conodonts and fusulinids of the Irgina Horizon of the Artinskian Stage (the *Neostreptognathodus* ex gr. *pequopensis* and *Pseudofusulina juresanensis–Parafusulina lutugini* zones).

At the Yuryuzan' River (Chuvashov *et al.*, 1990), radiolarians were found in the Upper Paleozoic section beginning with the Tastuba Horizon of the Sakmarian Stage. The radiolarian-bearing deposits of the section are as follows.

(1) The Sharipovo Formation spans the Tastuba and Sterlitamak horizons of the Sakmarian Stage, the fusulinid *Pseudofusulina moelleri* Zone, and the lower and middle parts of the fusulinid *P. urdalensis* Zone. The formation shows the alternation of mudstone, siltstone, and sandstone beds. In the lower part, there are predominantly mudstone packages, which include frequent concretions, interbeds, and lenses of pelitomorphic limestones. Some carbonate interbeds and concretions contain radiolarians. The total thickness of the formation is up to 800 m; that of the Sterlitamak beds varies from 200 to 250 m.

(2) The Urdala Formation (the upper part of the *P. urdalensis* Zone) is composed of pelitomorphic, frequently sponge-bearing limestones with rare interbeds of detrital limestones, marls, and mudstones. This unit is up to 75 m thick.

(3) The Bal'zyak Formation incorporates the Burtsevka Horizon of the Artinskian Stage and the fusulinid *Pseudofusulina concavutas–Pseudofusulina pedissequa* Zone. The lower part of the formation is mainly represented by mudstones and marls with interbeds and concretions of pelitomorphic limestones, frequently with very abundant radiolarians. Throughout the section, there are thin sandstone interbeds. The upper part of the formation shows members of flaggy marls (up to 7.5 m thick) rich in sponge, brachiopod, and fusulinid remains. The formation is 140–170 m thick.

In the lower part of the Bal'zyak Formation (the Yuryuzan' River site near the Musatovo Village), Amon identified radiolarians of the *Entactinosphaera crassiclathrata–Quinquerenis arundinea* Zone, such as *Entactinia* sp., *E. densissima* Naz. et Orm., *Entactinosphaera* sp., *Ent. crassiclathrata* Naz. et Orm., *Astroentactinia luxuria* Naz. et Orm., *Helioentactinia biexosphaera* Naz., *Copicyntra* spp., *C. cuspidata* Naz. et Orm., *C. phymatodonta* Naz. et Orm., *Copiellintra diploacantha* Naz. et Orm., *Tetracircunata* cf. *reconda* Naz. et Orm., and *Polyentactinia* sp. When examining radiolarians from the samples taken in the stratigraphic order, i.e., from the base upward, we revealed the following trend in morphological evolution of the assemblage. The samples of lower stratigraphic levels yielded the assemblage representing all morphological varieties of radiolarian skeletons, whereas the highest sample revealed the noticeable prevalence of genera *Copycintra* and *Copiellintra*, whose species have massive dense shells without long spines. Entactiniinae, Haplentactiniinae, and Latentisistulidae forms having delicate thin shells, long spines, and processes are abruptly reduced in abundance in the uppermost oryctoassemblage. This trend may reflect some environmental changes: either the temperature drop increasing the water density, or the basin shoaling that usually intensifies hydrodynamics; a combination of both factors is also possible.

(4) The Yanga-Tau Formation includes the Irgina Horizon of the Artinskian Stage (the fusulinid *Pseudofusulina juresanensis* Zone). The formation is represented by thick highly bituminous flaggy marls with limestone interbeds and members with fusulinid, sponge, brachiopod, and crinoid fauna. There are rare interbeds with abundant calcitized radiolarians. The

Lower Permian biostratigraphy of the southern Urals and Cis-Ural region

Stage	Horizon	Foraminiferal zone	Conodont zone	Zones and beds with radiolarians
Kungurian	Iren'		*Neostreptognathodus pnevi*	Beds with *Ruzhencevispungus uralicus*
	Filippovskoe			
Artinskian	Saranina			
	Sargina	*Parafusulina solidissima*	*Neostreptognathodus pequopensis*	*Polyentactinia lautitia*
	Irgina	*Pseudofusulina juresanensis–Parafusulina lutugini*		Beds with *Tetracircinata reconda*
	Burtsevka	*Pseudofusulina concavutas–Pseudofusulina pedissequa*	*Sweetognathus whitei*	*Entactinosphaera crassicalthrata–Quinqueremis arundinea*
Sakmarian	Sterlitamak	*Pseudofusulina urdalensis*	*Mesogondolella bisselli–Mesogondolella visibilis*	Beds with *Rectotormentum fornicatum*
				Camptoalatus monopterygius
				Entactinia pycnoclada–Tormentum circumfusum
	Tastuba	*Pseudofusulina verneuili–Pseudofusulina uralica*	*Mesogondolella lata*	*Helioentactinia ikka–Haplodiacanthus perforatus*
		Pseudofusulina moelleri	*Mesogondolella uralensis*	Beds with *Copicyntra sp./ Tetra-gregnon vimineum–Copiellintra diploacantha*
Asselian	Shikhany	*Sphaeroschwagerina sphaerica–Pseudofusulina firma*	*Mesogondolella pseudostriata*	
			Streptognathodus postfusus–Mesogondolella striata	Beds with *Tormentidae/ Haplo-diacanthus anfractus*
	Kholodnyi Log	*Sphaeroschwagerina moelleri–Pseudofusulina fecunda*	*Streptognathodus fusus–Mesogondolella simulata*	
			Streptognathodus constrictus–Mesogondolella adentata	
		Sphaeroschwagerina fusiformis	*Streptognathodus constructus–S. barskovi*	
			Streptognathodus sygmoidalis	
			Streptognathodus cristellaris	
		Daixina bosbytauensis–Daixina robusta	*Streptognathodus isolatus*	*Latentifistula crux*

formation is 130–170 m thick. The Yanga-Tau sediments undoubtedly represent a shallower facies than rocks of the Sharipovo Formation, and this corresponds to the trend of morphological evolution of the radiolarians upward the Bal'zyak section.

(5) The Gabdrashitovo Formation includes the Sargina and Sarana horizons of the Artinskian Stage (the *Pseudofusulina solidissima* and *Bairdia reussiana* zones). The formation is represented by regularly alternating sandstone and mudstone beds. There are interbeds of grainstones with fusulinids, bryozoans, brachiopods, crinoids, and algae. Rare mudstone interbeds of relatively considerable thickness (up to 40–50 cm) contain shells of ammonoids and radiolarians. Taxonomic composition of radiolarians from this stratigraphic level has not been studied yet, but affinity with the assemblage of the *Polyentactinia lautitia* Zone should not be excluded. The thickness is 95 m.

An almost analogous section is exposed northward up to the Ai River, but the Lower Permian radiolarians from this area have not been studied yet. Only radiolarians from the Kungurian deposits important for the regional biostratigraphy are investigated in detail. The Kungurian section in middle courses of the Ai River near the Alegasovo Village can be characterized as follows (Chuvashov *et al.*, 1990).

(1) The Filippovskoe Horizon–Taimeevo Formation: dark-gray mudstones, marls, and pelitomorphic

limestones. The apparent thickness is 3 or 4 m.

The Iren' Horizon of the Koshelevka Group.

(2) The Lemaza Formation:

(a) block- to coarse-clastic breccias with limestone clasts bearing fauna of the Artinskian age; the average size range of clasts is 15–20 cm; individual blocks are up to 1 or 2 m across; the thickness is up to 25 m;

(b) massive, thin or vaguely bedded, highly porous carbonate breccias; the predominant clast size range is 1–3 cm, rarely, 5–10 cm; the member is well exposed at the left bank of the Ai River near the Alegasovo Village; its incomplete thickness is 5 or 6 m;

(c) pale to yellowish gray flaggy marls up to 5 m in thickness.

(3) The Koshelevka Formation is represented by alternating siltstones, mudstones, and marls with subordinate thin interbeds of highly calcareous sandstones; the mudstone beds include numerous bun-shaped concretions of hard clayey limestone with radiolarians, ammonoids, and conodonts. The thickness is 12 m.

In addition to conodonts, Kozur (1980) described (from the Chuvashov's collection of 1978) new radiolarian genus and species *Ruzhencevisporites uralicus* Kozur from the lower part of this formation. Kungurian radiolarians were also found by Kozur (pers. commun.) from basal beds of the Koshelevka Formations in a section near the Aleksandrovkoe Village, 10 km to the north of the town of Krasnoufimsk (Fig. 1). At present, this is the northernmost known locality of the Early Permian radiolarians in the Uralian foredeep.

The deep-water part of the UF, which was the main habitat area of radiolarians, continues into the vast area of the deep-water Caspian depression (CD). The central part of the CD has not been adequately studied, whereas the marginal Upper Paleozoic deposits are recovered by many boreholes. The oldest Late Paleozoic radiolarians were found in the Serpukhovian deposits, and their representative assemblages are recovered from the mid-Asselian, Sakmarian, and Artinskian deposits of the Karachaganak reef area at the northern edge of the depression (Il'in, 1987; Afanas'eva, 1987; Afanas'eva *et al.*, 1986; Afanasieva and Zamilatskaya, 1993). Radiolarians are common in the Artinskian deposits in the western margins of the depression (Movshovich, 1977). They are also reported from the Upper Paleozoic deposits of the eastern Caspian depression. According to Afanas'eva *et al.* (1986), Borehole 11 in the Kenkiyak oil field recovered radiolarians similar with those from the Sarabil' Formation of the Ural region near Aktyubinsk. An identical assemblage was also found northward of the Kenkiyak oil field. Borehole 1 in the Aransai oil field yielded a rich radiolarian assemblage analogous to that from the Akstasy Formation at the Ural River. Radiolarians of the Artinskian age were also recovered by Borehole 3 in the Urekh-Tau oil field.

The generalized radiolarian zonation for the Lower Permian of the Southern Urals is presented in the Table. Radiolarian zones are correlated with zones of orthostratigraphic fossil groups.

LATE PALEOZOIC RADIOLARIAN PALEOBIOGEOGRAPHY OF THE URALS

Taphonomy of the Paleozoic radiolarians exhibits the sharp taxonomical and structural differences between their oryctocoenoses and taphocoenoses, and between the latter and assemblages of organisms inhabited sea and ocean basins in the remote geological past. In addition to the usual dissolution of skeletons of dead organisms in time and after their burial in bottom muds, radiolarian oryctocoenoses experienced dissolution, recrystallization (the primary organogenic opal is substituted by secondary opal or low-temperature quartz; in their turn, the latter can be substituted by calcite, apatite, fluorite, iron and manganese minerals, etc.), and mechanical deformation of skeletons during lithification, diagenesis, katagenesis, and subsequent metamorphism of sediments. For these reasons, the oryctocoenoses can hardly represent the real composition, diversity, and abundance of radiolarian assemblages and their paleoecological peculiarities. However, using the available data and the method of actualism (a comparison based on analogy), we can make some cautious paleogeographical and paleoecological interpretations and reveal some important trends.

Among the environmental factors controlling the geographic distribution, population density, and adaptation ability of radiolarians (i.e., finally, their facies confinement), the important are salinity, density, temperature and illumination of waters, basin depth, proximity or remoteness of habitat areas to or from the shoreline, hydrodynamics, presence or absence of currents and hydrographic barriers (Petrushevskaya, 1986; Nazarov, 1988; Kruglikova, 1990). As the main factors determining the distribution and abundance of radiolarians, we can indicate four of them (in order of the increasing influence): salinity, depth, hydrodynamics, and temperature.

Analysis of the radiolarian distribution in areas and facies zones of the Late Paleozoic basin of the Uralian foredeep, with due regard to the taphonomical and paleoecological peculiarities, led us to the following inferences (Fig. 5).

The radiolarian locality at the latitude of the town of Krasnoufimsk (approximately 57°N) marks the northern limit of their distribution in the UF. The occurrence of the monotaxon radiolarian assemblage, consisting of only one cool-water species *Ruzhencevispongus uralicus* Kozur and recovered from outcrops of the Koshelevka Formation (the Iren' Horizon of the Kungurian Stage) at the Alegasovo Village, Ai River, and Aleksandrovskoe Village near Krasnoufimsk, suggests unfavorable (cold-water) conditions. At this latitude,

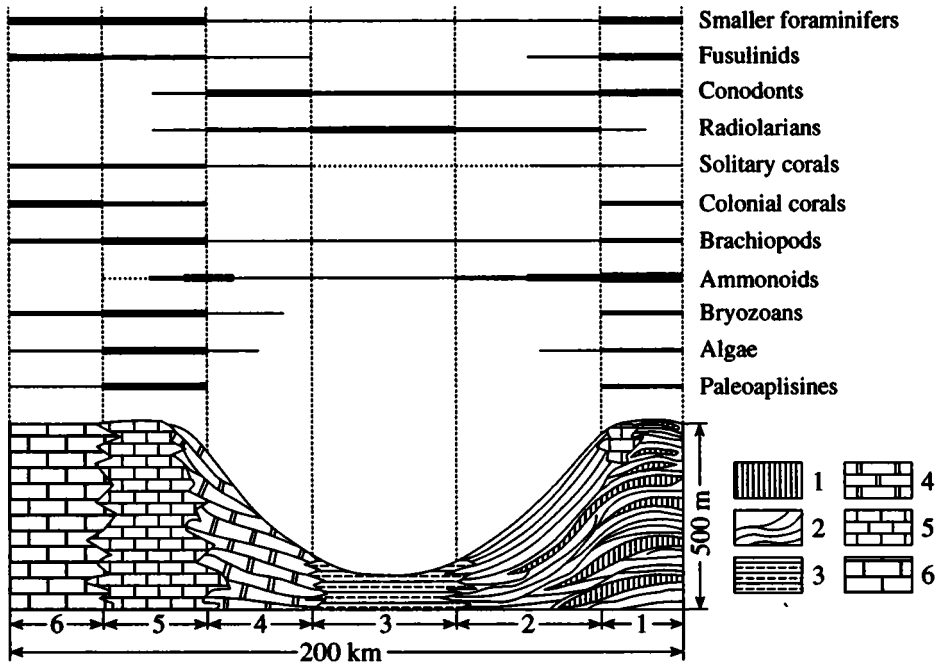

Fig. 5. Facial ranges of the main fossil groups in the Uralian foredeep (the late Asselian time): (1) coarse flysch; (2) fine flysch; (3) preflysch; (4) prereef; (5) reef; (6) bedded carbonates of the platform (out of scale).

the northern thermocline could be located, which served as the thermal barrier preventing the invasion of radiolarian biota into northern basins of the foredeep.

The southernmost basin of the UF inhabited by radiolarians (areas near towns of Orenburg and Aktyubinsk, as well as the Mugodzhary area, 50°–51°N), which belongs to the Tethyan region, was connected with the Caspian basin, especially in the Artinskian time (Afanas'eva *et al.*, 1986; Afanasieva and Zamilatskaya, 1993). In the Gzhelian–Artinskian time, this basin, spanning facies of the fine flysch, preflysch, prereef, and western biohermal (partly) zones, was inhabited by very abundant radiolarian assemblages of high taxonomic diversity. Such a degree of diversity and abundance of radiolarians (especially in the Gzhelian–Asselian time) is a unique phenomenon for the northern hemisphere (Nazarov, 1988). However, despite the basin connection with the Tethyan region, the radiolarian assemblages of the Gzhelian–Artinskian age cannot be attributed to the tropical realm, because they are of a mixed type and include forms representing shallow-water, cold-water, and cosmopolitan-tolerant morphotypes. This territory seems to have been a transitional zone from the Tethys to the boreal Peri-Tethyan basins. The abundance and diversity of radiolarian biota can be a result of favorable conditions in terms of salinity, depth, hydrodynamics, and temperature, as well as of plentiful biogenic nutrients (autotrophs, phosphorus, silica, nitrites, and oxygen) transported from neighboring areas of the eastern edge of the foredeep and its northern part.

At the latitude of the interfluve between the Ural and Sakmara rivers, there probably was another thermal barrier strongly affecting the northward dispersal of radiolarians, although not so critical as the thermocline at the latitude of Krasnoufimsk. This barrier produced a selective affect: it closed the way to the north for some assemblages or caused changes in their composition, but it did not prevent the distribution of others.

For instance, in the sections to the north of the Ural and Sakmara rivers interfluve, we noted the following distribution peculiarities of radiolarians:

(a) the Gzhelian *Tormentum pervagatum* and *Tormentum protei* zones, the Asselian *Latentifistula crux*, Tormentidae, and *Copicyntra* sp. zones; the Sakmarian *Camptoalatus monopterigus* Zone; and the Artinskian *Rectotormentum fornicatum* Zone have not been recognized here;

(b) assemblages of the Gzhelian *Tormentum pervagatum* and *Tormentum protei* zones (correlatives with the *Tormentum pervagatum* and *Tormentum protei* beds in the Bala-Tau section at the Belaya River in the preflysch zone) show considerable taxonomic changes;

(c) the following units established within the area under discussion: the Sakmarian *Helioentactinia ikka–Haplodiacanthus perforatus* Zone in the fine flysch facies of the Tastuba Horizon of the Sim section; the Sakmarian *Entactinia pycnoclada–Tormentum circumfusum* Zone in the fine flysch facies of the Sterlitamak Horizon of the Ishtuganovo section at the Belaya River, in the preflysch facies of the Tyul'kas section and in the fine flysch facies of the Sim section; the Artinskian

Quinqueremis arundinea Zone in the fine flysch facies of the Burtsevka Horizon of the Ishtuganovo section, in the fine flysch facies of the Sim section, and in the preflysch facies of the Tyul'kas section; the Artinskian *Tetracircinata reconda* Zone in the Sargina Horizon and fine flysch facies of the northern edge of the Sim depression; and the Artinskian *Polyentactinia lautitia* Zone in the fine flysch facies of the Sargina and Sarana horizons of the Siryat' section at the Belaya River. Finally, to the north of the interfluve between the Ural and Sakmara rivers, we distinguished some new cold-water radiolarian assemblages, e.g., those characteristic of *Pseudoalbaillella* Beds (the Asselian Stage of the Usol'ka section, preflysch facies) and *Ruzhencevispongus uralicus* Beds (fine flysch facies of the Kungurian Stage, the Iren' Horizon of the Alegazovo section at the Ai River and of the Aleksandrovskoe section in the Krasnoufimsk area).

Evidently, radiolarians did not inhabit the upper littoral area corresponding to the facies zone of the coarse flysch. Because of small depth, active hydrodynamics, decreased salinity, and unfavorable conditions of shell burial, radiolarians were either absent or so scarce there that they left no paleontological records. However, radiolarians were able to rapidly migrate through the calm-water gulfs and bays near the shore, as demonstrated by the Shafeevka River section.

The situation was similar in the facies zone of biogenic buildups. The unfavorable conditions of increased salinity and small depth hampered the dispersal of radiolarians behind the typical reefs. Situation in the frontal part of reefs facing the open basin was slightly different. Exceptionally interesting data of Afanas'eva on the occurrence of radiolarians in the lagoonal facies of the Karachaganak atoll (the northeastern Caspian region) (Afanas'eva, 1986; Afanas'eva *et al.*, 1986; Afanasieva and Zamilatskaya, 1993) and our observations in the Yurak-Tau massif show that radiolarian communities could inhabit the frontal part of reefs and the prereef zone. It should be noted that some radiolarian assemblages from the Middle Carboniferous (Bashkirian) and Lower Carboniferous (Serpukhovian) reef deposits of the northern Caspian region (Afanas'eva, 1986; Afanas'eva *et al.*, 1986) are dominated by sedentary forms of the Pylentonemidae family, which are more resistant to environmental changes than planktonic radiolarians.

CONCLUSION

The presented data on biostratigraphic and facies distribution of the Late Carboniferous–Early Permian radiolarians in the Uralian foredeep allows us to make the following conclusions:

(1) In the Upper Paleozoic deposits of the Cis-Urals, radiolarians appeared immediately after formation of a relatively deep-water basin characterized first by clayey–siliceous and then by sandy–clayey flysch sedimentation. The radiolarian distribution was limited in the east by the nearshore zone, whose fragments are preserved until the present time. The western limit of the radiolarian distribution coincided with the linear chain of organogenic buildups till the Artinskian time inclusive and by the high-salinity zone in the Kungurian time.

(2) Along the eastern and western limits, radiolarians show discrete distribution throughout the section and are frequently missing in large stratigraphic intervals. The sections with abundant radiolarians are located at the eastern slope of the Uralian foredeep (the fine flysch zone, Figs. 3 and 5) and in the depression zone. The latter is exemplified by the Krasnousol'sk section, where radiolarians occur in the interval from the Upper Carboniferous to the Artinskian Stage. The data on aforementioned sections and marginal areas of radiolarian distribution can serve as a basis for elaborating a valid radiolarian zonation in the territory under consideration in order to use it instead the current stratigraphic scheme of "radiolarian beds."

(3) Radiolarians were found only in the southern part of the Cis-Ural basin, and this limitation may be explained by the existence of an ecological barrier preventing the northward migration of these organisms. The barrier most probably corresponded to a zone of the temperature drop in the basin. Analysis of the Early Permian biota as a whole shows a decrease in the diversity and abundance of many faunal groups in the northern direction (Chuvashov, 1991). This trend well accords with the replacement of the Kungurian evaporites and red deposits by coal-bearing formations in the same direction.

(4) The radiolarian zonation in the southern Urals can provide an effective base for correlation between the Lower Permian sections of the stratotype area and the Tethyan region.

ACKNOWLEDGMENTS

This work was supported by the Russian Foundation for Basic Research, project no. 94-05-17718, and by the Peri-Tethys Programme, project no. 95-62.

Reviewers A.S. Alekseev,
I.A. Basov, and V.S. Vishnevskaya

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Stratigraphy and Geological Correlation, Vol. 7, No. 1, 1999, pp. 49–58. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 7, No. 1, 1999, pp. 56–65.
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Reference Section and Upper Permian Zonation in Southeastern Mongolia

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Received June 27, 1997

Abstract—A reference section of the Upper Permian deposits of the Tethyan basin in southeastern Mongolia is defined. The brachiopod zonation of the deposits is described. Two zones corresponding to one transgressive phase are united into a horizon. The distribution of brachiopod zones within the areas of Upper Permian exposures in different structural-facial zones is outlined. Brachiopod assemblages of the zones are correlated with those from coeval units in the Boreal basin of northeastern Mongolia, Arctic regions, and central and northeastern Asia. Late Permian migration paths of brachiopods of the Ussuri Province are considered. Available paleontological records point to the existence of the Late Paleozoic South Mongolian microcontinent, whose shelf was inhabited by peculiar fauna of the mixed Boreal–Tethyan type.

Key words: *Mongolia, Upper Permian, reference section, zonation, migration, zoogeography.*

Two marine basins of the Permian time are known in Mongolia. The Khangai–Khentei basin of the Mongolia–Transbaikal area in the Boreal zoogeographic province occupied the northeastern and central parts of the country. The South Mongolian basin in the southeastern areas was a part of the Ussuri zone in the northern margin of the Paleotethys. The basins were disconnected in the Mongolian territory and characterized by few common faunal elements. Meanwhile, both basins reveal some common features in their evolution and faunal change, which can be attributed to the same succession of abiotic events discussed below.

In southeastern Mongolia, the Upper Permian deposits are exposed in a relatively narrow latitudinal band about 800 km long and up to 150 km wide. Their thickness locally exceeds 5 km. Marine deposits, predominantly terrigenous sediments and limestones, occur in four structural-facial zones established by O.B. Suetenko (see Pavlova *et al.*, 1991) and replacing one another from the south to the north. These are the Solonker, Dalan-Ula–Lugin-Gol, Totoshan', and Khubsugul–Bairim-Obo zones (Fig. 1). The deposits contain diverse fossils, among which foraminifers, brachiopods, and bryozoans are most valuable for zoogeographic and biostratigraphic reconstructions.

In the monograph "*Permian Invertebrates of Southern Mongolia*," Pavlova *et al.* (1991) described seven faunal assemblages. Four pre-Kungurian (Early Permian) assemblages are dominated by foraminifers, and three younger ones, by brachiopods and bryozoans. However, the localities studied are mostly confined to thin unpersistent beds or isolated exposures, and the extension of fauna-bearing beds along the strike and their succession in a single section have not been followed.

The Upper Permian section described below is more than 450 m thick and characterized by successive assemblages of brachiopods and other fossils. Some components of these assemblages were recognized earlier in other sections of different structural–lithologic zones (Fig. 1). The section under consideration shows successive occurrences of all assemblages including the youngest one unknown previously from Mongolia. On this basis, the section is assumed to be the reference one and appropriate for recognizing the biostratigraphic zonation in its deposits.

The section is located at the southern boundary of the Dalan-Ula–Lugin-Gol structural-facial zone (Fig. 1, Site 6), 1.5 km to the northwest of the summit of the Dzhirem-Ula Mountain (Fig. 2).

In the 1960s and 1970s, scarce brachiopods and bryozoans were collected from the Upper Permian deposits of the Dzhirem-Ula area by B.M. Kazakov, O.D. Suetenko, S.S. Lazarev, and myself (Pavlova *et al.*, 1991). The bed succession and fossils of the section were described by T.A. Grunt, E.Ya. Leven, and myself in 1990 and 1994. The descriptions are used as a basis for this article.

There are different interpretations of the Upper Permian sequence position and structure of the section due to previously insufficient data on geology of the region complicated in tectonic patterns. In the geological map, scale 1 : 500000 (compiled by A.F. Boishenko and I.P. Symev in 1984), the Upper Permian beds of the latitudinal extension are shown to have tectonic contacts with Cretaceous deposits in the north and with undi-

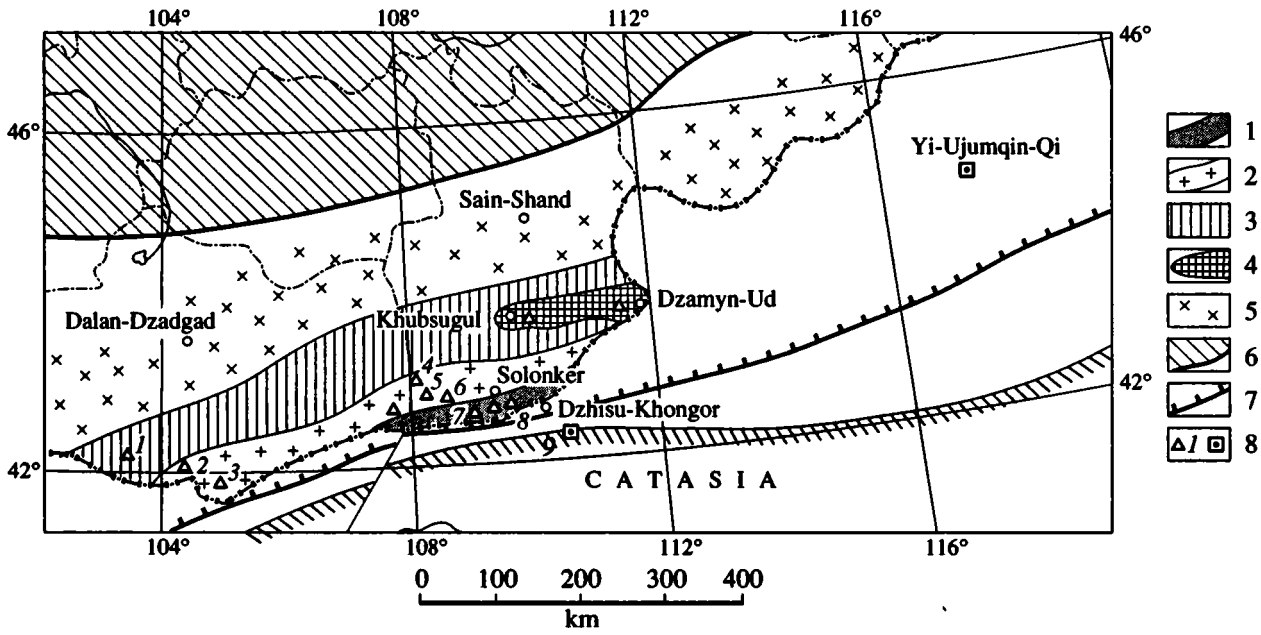

Fig. 1. Tectonic scheme of Mongolia with indication of biogeographic elements and faunal localities.
(1–4) structural-facial zones: (1) Solonker; (2) Dalan-Ula–Lugin-Gol; (3) Totoshan'; (4) Khubsugul–Beirim-Obo; (5) South Mongolian microcontinent; (6) Siberian (Salair) continent; (7) dividing line between the Siberian and Catasian plates (7 and 8 according to Ruzhentsev *et al.*, 1992); (8) major Permian faunal localities, sites in the areas of (*1*) Ulan-Shivenyi-Khuduk Well; (*2*) Khuryin-Chibe-Khuduk Well; (*3*) Mungun-Obo Mountain; (*4*) Tumen-Ul'dzei-Ula Mountain; (*5*) Lugin-Gol River; (*6*) Dzirem-Ula Mountain; (*7*) Onts-Obo Mountain; (*8*) Khoyar-Khuduk Well; (*9*) Khara-Erdene-Ula.

vided Middle Carboniferous–Lower Permian deposits in the south.

According to another map, scale 1 : 100000 (compiled by G. Badarch and Ya. Davga-Ochir in 1995; Fig. 2), the same beds are shown in contacts with green micaceous shales (presumably Devonian) and locally with acid extrusive rocks in the north, and as overlain by the Permian acid rhyolites in the south.

When working with E.Ya. Leven in the field in 1990, I shared his opinion on the overturned position and tectonic contact between the beds in the north. The overturned position was mainly inferred from the occurrence of abundant small brachiopods near the northern edge of the valley, which we took in the field for *Echinauris jisuensis* (Chao). This species was found in the upper beds dated back to the Midian in a section located in the southeast, within the Dzhisu-Khongor area of northern China. The subsequent comprehensive investigation of brachiopods and their successive assemblages allowed me to establish the normal bedding in the section under consideration.

First, being prepared and analyzed, the small brachiopods from the northern edge of the valley (i.e., from the basal beds of the section in the normal position) appeared to represent a new species *Anemonaria sulankherensis* similar to the species I described earlier from the basal Upper Permian beds of southeastern Mongolia.

Second, the middle part of the section yielded fossils widespread in many Ufimian–Kazanian sections of four aforementioned structural–lithologic zones.

Third, the beds of the southern edge of the valley (the Uppermost Permian terrigenous sediments overlain by rhyolites) contain the youngest brachiopod assemblage of the Late Permian, including the mentioned *Echinauris jisuensis* previously unknown from southeastern Mongolia.

My field investigation of 1994 and collection of fossils from the section (geological map of the area was compiled by the Mongolian Geological Survey at that time, Fig. 2) confirmed the normal position of beds in the section and supported the suggestion of Boishenko and Syrnev about the tectonic contacts at the base and top of the Upper Permian Lugin-Gol Formation.

In the area under consideration, the Upper Permian Lugin-Gol Formation is represented by alternating beds of polymictic sandstones of variable grain size, gritstones, calcareous siltstones, and limestones, all striking in a sublatitudinal direction (dip azimuth 140°–150°, angle 45°–60°). Abundance of carbonate fine-grained deposits increases upsection (Fig. 3). To the north of the Upper Permian exposures, there are heavily folded metamorphic micaceous greenschists supposedly Devonian in age and also striking in a sublatitudinal direction. Both units may be separated by tectonic contact. A thick (350–400 m) sequence of acid extrusive rocks rests above different Upper Permian beds,

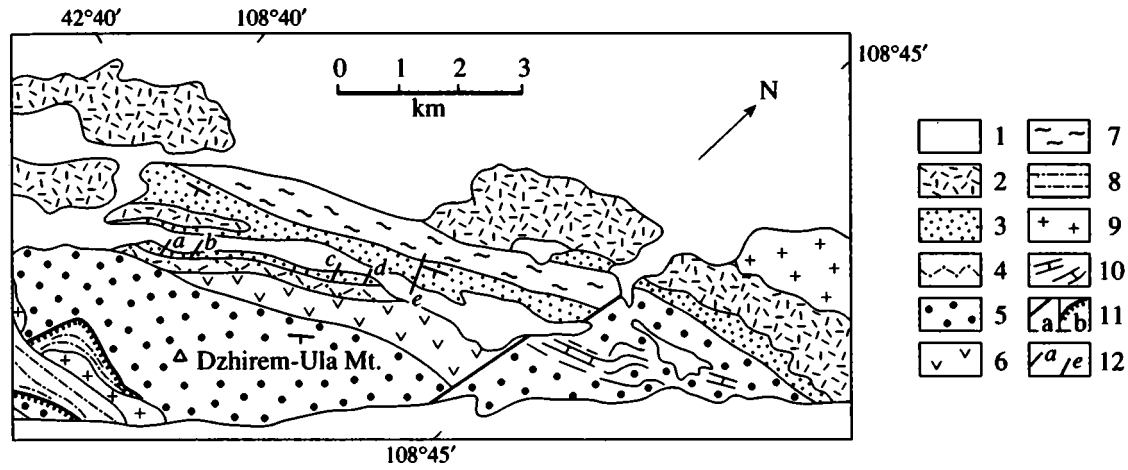

Fig. 2. Geological map of the Dzhirem-Ula Mountain area (compiled by G. Badarch and Ya. Dagva-Ochir in 1995).
(1) Quaternary deposits; (2) acid extrusive rock; (3) Permian sandstone, siltstone, and limestone; (4) rhyolites and their tuffs (P); (5) arkose, sandstone (C); (6) andesites and their tuffs (C); (7) micaceous shale (D?); (8) metasandstone, siltstone; (9) granite; (10) limestone lens; (11) fault: (a) steeply dipping; (b) thrust; (12*a*–12*e*) section line.

which are locally baked (the color of calcareous siltstones is changed from dark-gray to pink-red). The Upper Permian deposits are well exposed. Only in the middle of the valley (Bed 3 of the section), they are covered by thin eluvial sediments locally enclosing dense sandy beds. The faunal remains consist of brachiopods, bryozoans, bivalves, corals, and crinoids. Of these fossils, brachiopods are most abundant, diverse, polyfacial, and valuable in stratigraphic aspect. They are used to reveal the biozonation of the section.

Table 1 shows the brachiopod ranges in the section and three distinct changes in their assemblages. Taking into consideration the brachiopod distribution in other sections, I distinguish three units corresponding from the base upward to the *L. decimana–W. angustata*, *A. lita–K. adressum*, and *E. jisuensis* brachiopod zones.

Liosotella decimana–Waagenoconcha angustata Zone.

Index species: *Liosotella decimana* Man., 1991, and *Waagenoconcha angustata* Sar., 1984.

Type area. In the reference section, the zone is represented by the 380-m-thick unit of polymictic, predominantly coarse-grained calcareous sandstones alternating with gritstones and slitstones (Beds 1–4, Fig. 3).

Zonal assemblage in the reference section (Table 1). Brachiopods *Arctitreta kempei* (Anders.), *Orthotetes jugorica* Ustr., and *Cancrinella koninckiana* (Keus.) characteristic of the zone are also known from the "Khovs-Gol" assemblage of the Kungurian age (Pavlova *et al.*, 1991). In addition to these and index species, *Anemonaria sulankherensis* Man., *Waagenoconcha impercepta* Man., and *Yakovlevia mammatiformis* (Frks.) occur only within the zone. *Yakovlevia kaluzinensis* Frks., *Kochiproductus saranaeanus* (Frks.), *Kaninospirifer* cf. *insertiplicatus* Pavl., and *Spiriferella keilhaviiformis* Frks. pass into the overlying zone.

The zone also incorporates bryozoans *Maychella diserta* Gorjunova et Morozova, *M. nitens* Gorjunova et Morozova, *Maychellina aliena* Goryunova et Morozova, *Permofenestella labuensis* (Morozova), and *P. vera* Morozova.

Distribution. At the distance of 400 km to the west, in the Khuryin-Chibe-Khuduk Well area of the southwestern flank of the Dalan-Ula–Lugin-Gol structural–lithologic zone (Fig. 1, Site 2), the zonal assemblage was found in the lower Lugin-Gol Subformation 1500 m thick. Abundant brachiopods are confined to the 120-m-thick unit (Beds 1–8) of alternating conglomerates, gritstones, and sandstones of variable grain size (Pavlova *et al.*, 1991).

These deposits contain brachiopod species known from the reference section: *A. kempei, L. decimana, Y. mammatiformis, K. saranaeanus, W. impercepta, K. insertiplicatus*, and *S. keilhaviiformis*; those of zonal significance, but missing in the reference section: *Chonetinella sinuata* Afan. and *Stenoscisma timorensis* (Hayasaka et Gan). The brachiopods coexist here with bivalves *Phestia* sp. and bryozoans *Polypora principalis* Gorjunova and *Maychillina aliena* Gorjunova et Morozova.

In the reference section, the Lugin-Gol Formation contacts the Devonian deposits, whereas in the southwestern flank of the Dalan-Ula–Lugin-Gol structural–lithologic zone, 2.9 km to the north of the Dzil'-Khailiste-Khuduk Well, it lies conformably on the Kungurian(?) sandy–shaly Bulgan-Ula Formation 2000 m thick. Faunal remains were found in the upper part of the formation consisting of intercalated flyshoid conglomerates, sandstones, and siltstones. *Kolymaella ogonorensis* (Zavod.) found among brachiopods, which are not well preserved, is the index species of the lower unit of the Omolon Horizon (Anonymous, 1990),

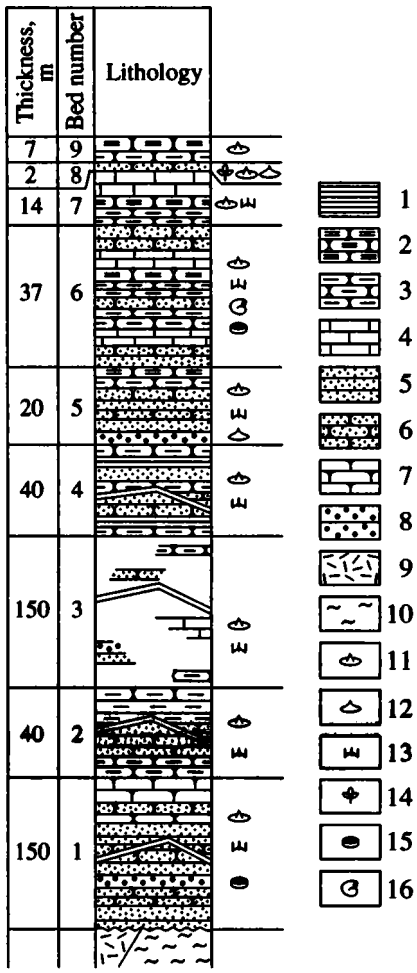

Fig. 3. Upper Permian reference section in the Dzhirem-Ula Mountain area (scale 1 : 1000).

(1) Mudstone; (2) calcareous mudstone; (3) calcareous siltstone; (4) limestone; (5) sandstone; (6) calcareous sandstone; (7) detrital limestone; (8) gritstone; (9) acid extrusive rock; (10) micaceous shale; (11) brachiopods; (12) bivalves; (13) bryozoans; (14) flora; (15) crinoid columnals; (16) smaller foraminifers.

which is dated as Ufimian in age. Bryozoans are represented by thick-branched colonies of *Iraidina maxima* Morozova also characteristic of the Ufimian deposits. However, the associated ammonoid species *Globioceras elenae* Bogosl. may indicate the Kungurian age of the unit according to the development state of its lobe line (Pavlova *et al.*, 1991).

In the Mungun-Obo Mountain area (Fig. 1, Point 3), the zone is represented by the 90-m-thick member of alternating conglomerates, sandstones, and organogenic limestones, which are typical of the lower part of the Lugin-Gol Formation up to 3000 m thick in this area. The deposits yielded brachiopod species known from the reference section: *L. decimana, A. kempei, O. jugorica, Y. mammatiformis*, and *S.* aff. *keilhaviiformis*, as well as other species of the zonal assemblage: *Spiriferella mugunica* Pavl. and *Spirelytha* sp.

In the isolated exposure of the Lugin-Gol Formation, 15.5 km from the Khadtyn-Bulak Spring at the Lugin-Gol River (Fig. 1, Site 5), the zone contains brachiopods *W. angustata, O. jugorica, Y. mammatiformis*, and *S.* aff. *keilhaviiformis*, as well as species missing in the reference section, such as *Anemonaria minuskula* (Man.), *Kochiproductus maximus* Man., *Megousia* sp., and *Krotovia* sp.

Age. The underlying Bulgan-Ula Formation is characterized by faunal remains of the Kungurian age. The index-species *Waagenoconcha angustata* is recorded in the Starostin (the Ufimian Stage) and Selander (the Kazanian Stage) formations of Spitsbergen and the Lek-Vorkuta Formation (the Kungurian and Ufimian stages) of the Pai-Khoi.[1] The index species *Liosotella decimana* was identified as *L. spitzbergeniana* (Toula) from the Zhesi Formation (the Ufimian and Kazanian stages) in the Bei-Shan (Tsyan Chun-Fa *et al.*, 1963) and Xi-Ujumqin-Qi regions (Lee Li *et al.*, 1982, 1983) of northern China. Amid species occurring only within the zone, *Orthotetes jugorica* is known from the Starostin Formation of Spitsbergen, the Lek-Vorkuta Formation of the Pai-Khoi, and the Kozhim Rudnik Formation (the Ufimian Stage) of the Urals. *Waagenoconcha impercepta* is typical of the Lek-Vorkuta Formation of the Pai-Khoi, and *Yakovlevia mammatiformis* marks the upper Artinskian–Ufimian beds of the western Arctic sector.

Brachiopods passing into the overlying zone are *Spiriferella keilhaviiformis* known from the Vladivostok Horizon (Ufimian and Kazanian stages) of the southern Primor'e and from the lower Kanokura Formation (Midian Stage) of Japan, and *Yakovlevia kaluzinensis* was found in the upper Lower Permian beds of the Ussuri region and in the Chandalaz Horizon (Midian Stage) of the southern Primor'e region (Likharev and Kotlyar, 1978).

Bryozoans are represented by the species characteristic of Mongolia, except for *Permofenestella labuensis* occurring in the Ufimian deposits of the Omolon Horizon of northeastern Russia and in the lower Chandalaz beds of the southern Primor'e region.

Because the zonal assemblage contains species, whose age ranges from the Early Permian to the Ufimian and also from the Ufimian to younger levels, the zone can be attributed to the Ufimian.

Alispiriferella lita–Kaninospirifer adressum Zone.

Index species: *Alispiriferella lita* (Fredericks, 1924), and *Kaninospirifer adressum* (Lui et Waterhouse, 1985).

[1] Stage ranges of formations and horizons correspond to those accepted in *Zonal'naya* ..., 1991; Kalashnikov, 1993; Anonymous, 1990; Manankov, 1998; Pavlova *et al.*, 1991; *Pozdnepermskii* ..., 1984; Tsyan Chun'-Fa *et al.*, 1963; Ding Yunjie *et al.*, 1985; Grabau, 1931; Lee Li *et al.*, 1982, 1983.

Table 1. Brachiopod ranges in the reference section

Species	Bed number								
	1	2	3	4	5	6	7	8	9
	Thickness, m								
	150	40	150	40	20	37	14	2	7
Echinauris jisuensis (Chao)							—	—	—
Paramarginifera peregrina Frks.							—	—	—
Neospirifer yihewusuensis Duan et Li							—	- -	—
Marginifera leptorugosa Duan et Li									—
Cleiothyridina ex gr. *subexpansa* Waag.									—
Strophalosia kharaerdensis Man.								—	
Anidanthus ussuricus (Frks.)							—	—	
Horridonia morrisi (Chao)							—	—	
Streptorhunchus sp.							—	—	
Streptorhunchus broilii Grabau							—		
Leptodus ex gr. *nobilis* (Waagen)							—		
Stenoscisma purdoniformis (Grabau)							—		
Windhamia tjuvaevi Man.							—	—	
Mongolochonetes inaequalis Afan.							—	—	
Marginifera gobiensis Chao					—	- -	- -	—	—
Poikilosakos cf. *kamiyassensis* Tazawa et Takaizumi					—	- -	—		
Waagenoconcha sp.					—	- -	—		
Megousia cf. *sinuata* (Ustr.)					—				
Globiella reshetovi Man.					—	—			
Alispiriferella lita (Frks.)					—	—			
Kaninospirifer adressum (Lui et Water.)					—	—			
Yakovlevia dzhiremulensis Man.					—	- -	—	—	
Yakovlevia kaluzinensis Frks.				—	—	- -	- -	—	—
Kaninospirifer cf. *incertiplicatus* Pavl.				—	—	—			
Spiriferella keilhaviiformis Frks.				—	—				
Arctitreta kempei (Andersson)		—	- -	—	- -	—			
Kochiproductus saranaeanus (Frks.)	—	—	- -	—	- -	—			
Cancrinella koninckiana (Keys.)	—								
Yakovlevia mammatiformis (Frks.)		—	- -	—					
Waagenoconcha angustata Saryts.		—							
Waagenoconcha impercepta Man.	—	—	- -	—					
Liosotella decimana Man.	—	—	- -	—					
Orthotetes jugorica Ustr.	—	—							
Anemonaria sulankherensis Man.	—								

Type area. In the reference section, the zone 57 m thick is predominantly represented by calcareous siltstones of variable grain size, mudstones, and limestones of the Lugin-Gol Formation (Fig. 3, Beds 5 and 6). The unit conformably overlies the *L. decimana–W. angustata* Zone. The lower boundary of the zone is placed at the last occurrence level of abundant brachiopods *Anemonaria* and *Liosotella*, which is simultaneously the first occurrence level of index species and *Globiella reshetovi*, the latter widespread in coeval deposits of the southernmost Solonker structural–lithologic zone. The zonal assemblage includes abundant and diverse bryozoans.

Zonal assemblage of the reference section. Brachiopods *Artitreta kempei* (Anders.), *Yakovlevia kaluzinen-*

sis Frks., *Kochiproductus saranaeanus* (Frks.), *Spiriferella keilhaviiformis* Frks., and *Kaninospirifer* cf. *incertiplicatus* Pavl. pass into the unit from the underlying zone. In addition to index species, *Megousia sinuata* (Ustr.) and *Globiella reshetovi* Man. are typical of the zone. *Mongolochonetes inaequalis* Afan., *Marginifera gobiensis* Chao, *Yakovlevia dzhiremulensis* Man. sp. nov., *Y. kaluzinensis* Frks., *Poikilosakos* cf. *kamiyassensis* Tazawa et Takaiz., and *Waagenoconcha* sp. pass into the overlying zone. Coexisting bryozoans are *Dyscritella turbini* Romantchuk, *D. praespinigera* Romantchuk, *Permofenestella labuensis* (Morozova), and *Polyporella lyndoni* (Ph. Ross).

Distribution. In the exposure at the Khuryin-Chibe-Khuduk Well (Fig. 1, Site 2), the zone is represented by the 78-m-thick member of sandstones of variable grain size alternating with gritstones and conglomerates (Beds 9–15, Fig. 4, Pavlova *et al.*, 1991). The upper part of the member includes organogenic limestones with abundant faunal remains. Brachiopods missing in the reference section are *Chonetinella sinuata* Afan., *Stenoscisma timorensis* Hajasaka et Gan, and *Kaninospirifer striatoplicatus* (Gobbet). Associated bryozoans are identified as *Cyclotrypa admiranda* Morozova, *Dyscritella fida* Morozova, *D. turbini* Romantchuk, *Fistulipora egregia* Morozova, *Maychellina aliena* Gorjunova et Morozova, *Minilya paratuberculifera* Yang et Loo, *Permofenestella simplex* Morozova, *Poliporella lyndoni* (Ph. Ross), *Ramiporidra ramificata* Morozova, *Rectifenestela gilianshanensis* (Yang et Loo), *R. ornatiformis* Gorjunova, and *R. originaria* Morozova. Macrofauna coexists with smaller foraminifers *Calcitornella* ex gr. *stephensoni* (Howchin), *Endothyra* sp., and *Nodosaria* cf. *kolymica* M.-Maclay and with bivalves *Cypricardinia permica* Licharew.

In the isolated exposure of the Lugin-Gol Formation, 1.5 km to the northeast of the Ulan-Shivenii-Khuduk Well (Fig. 1, Site 1), S.V. Ruzhentsev found brachiopods *Liosotella* sp., *Waagenoconcha* sp., *Yakovlevia* sp., *Stenoscisma timorensis* Hayasaka et Gan, *Alispiriferella lita* (Frks.), *Kaninospirifer striatoparadoxus* (Toula), and *Spiriferella saranaeformis* (Frks.), which indicate the correspondence of the enclosing deposits to two zones described above.

Age. The index species *Alispiriferella lita* is known from the Vladivostok Horizon of the southern Primor'e region, the Kanokura Formation of Japan, and the *Magadania bajkurica* local zone (the Kazanian Stage) of the Omolon massif. The index species *Kaninospirifer adressum* is recorded in the Zhesi Formation of northern China. The stratigraphically significant species *Globiella reshetovi* is characteristic of the Kazanian deposits of the East European platform and the Kanin Peninsula. Representatives of the same genus are known from the Lower Permian deposits of Australia, Afghanistan, and southeastern Pamirs. *Megousia sinuata* occurs in the Zhesi Formations of Bei-Shan and Xi-Ujumqin-Qi regions of northern China. *Yakovlevia dzhiremulensis* sp. nov., which passes into the succeeding assemblage, was described under different names from the Zhesi Formation of the Bei-Shan region. *Marginifera gobiensis* shows the same distribution pattern. *Poikilosakos kamiyassensis* was described from the Kanokura Formation of Japan.

Bryozoans *Discritella turbini* and *D. praespinigera* are known from the Osakhtinskaya Formation (Kazanian Stage) of the Khabarovsk region and from the Omolon Horizon of northeastern Russia. *Permofenestella labuensis* and *Polyporella lyndoni* were found in the Omolon Horizon of northeastern Russia and the Chandalaz Horizon of the southern Primor'e region.

Foraminifer species *Nodosaria kolymica* is known from the Omolon Horizon.

The above-mentioned facts allow the zone to be attributed to the Kazanian.

The deposits bearing fossil assemblages of two zones described above are united into the Tsagan-Ula Horizon (Table 2), whose stratotype corresponds to the zone stratotypes in the reference section at the Dzhirem-Ula Mountain. The respective large area of distribution of marine sediments suggests an extensive cool-water transgression of the Permian time in Mongolia and some other regions of Asia. The transgression also left behind the Ufimian–Kazanian deposits of the Tsagan-Temete Horizon of central and northeastern Mongolia (Manankov, 1995), which bears the boreal assemblage similar to those of the Omolon Horizon of the Kolyma–Omolon massif and also of the Zhesi Formation of northern China, where, like in southeastern Mongolia, the West Arctic elements prevail amid fossils.

Echinauris jisnuensis Zone of the Solonker Horizon

Index species: *Echinauris jisuensis* (Chao, 1927).

Type area. In the reference section, the zone is represented by the 23-m-thick member of alternating limestones, calcareous mudstones, siltstones, and sandstones (Beds 7–9) overlying conformably the *A. lita–K. adressum* Zone. The zone base is defined at the disappearance level of some taxa of the preceding assemblage, such as *Arctitreta, Kochiproductus, Alispiriferella,* and *Kaninospirifer adressum.* Representatives of the genus *Strophalosia* that replaced the genus *Aulosteges* in Siberian sea basins of the Kazanian–Tatarian time (Kalashnikov, 1993) appeared and became widespread above this level. Among bryozoans, there appeared scarce species of genera *Meekopora, Hexagonella,* and *Discritellina.*

Zonal assemblage of the reference section. Brachiopods of the underlying zone are represented in the assemblage by *Mongolochonetes inaeqalis* Afan., *Marginifera gobiensis* Chao, *Yakovleva kaluzinensis* Frks., *Y. dzhiremulensis* Man., *Poikilosakos kamiyassensis* Tazawa et Kakaisumi, and *Waagenoconcha* sp. Brachiopods new-comers are *Echinauris jisuensis* (Chao), *Streptorhunchus broilii* Grabau, *Streptorhunchus* sp., *Marginifera leptorugosa* Duan et Li, *Paramarginifera*

peregrina Frks., *Anidanthus ussuricus* (Frks.), *Horridonia morrisi* (Chao), *Strophalosia kharaerdensis* Man., *Leptodus* ex gr. *nobilis* (Waag.), *Windhamia tjuvaevi* Man., *Waagenoconcha* sp., *Martinia* sp., *Cleiothyridina* ex gr. *subexpansa* Waag., and *Neospirifer yihewusuensis* Duan et Li.

Bryozoans of the zone are *Discritellina grandiscopa* Gorjunova, *Maychella diserta* Girjunova et Morozova, *Meekopora venusta* Gorjunova et Morozova, *Permofenestella labuensis* Morozova, *Rectifenestella totoshaensis* (Morozova), and *Streblascopora confusa* Morozova.

Distribution. In the Dalan-Ula–Lugin-Gol structural–lithologic zone, 20 km to the south of the Tumen-Uldzei-Ula Mountain (Fig. 1, Site 4), Suetenko, who studied the Lugin-Gol deposits, identified an impoverished zonal assemblage that includes brachiopods *Anidanthus ussuricus*, *Marginifera* cf. *gobiensis*, *Waagenoconcha nalivkini* Frks., and *Neospirifer* sp. in association with bryozoan species *Maychellina ornata* (Morozova).

In the southernmost Solonker structural–lithologic zone, the brachiopod assemblage occurs in the Tavan-Khara Formation. The 1200-m-thick formation is composed of andesitic and andesitic–dacitic porphyrites, and liparitic porphyries and their tuffs locally intercalated with sandstone beds and limestone lenses. In the Onts-Obo Mountain area (Fig. 1, Site 7), the 5-m-thick calcareous sandstone member in the middle of the Tavan-Khara Formation yielded brachiopods indicating that the host deposits correspond to the boundary beds between the *A. lita–K. adressum* and *E. jisuensis* local zones. The brachiopod assemblage includes *Echinauris jisuensis, Marginifera gobiensis, M. leptorugosa, Globiella reshetovi, Anidanthus spineus* Man., and Terrakea echinata Man. It is significant for biogeographical reconstructions that the two last species were first described from the Ufimian–Kazanian deposits of the Boreal Khangai–Khentei basin (Manankov, 1992) in the Mongolian–Transbaikal region of northeastern Mongolia.

In the Khara-Erdene-Ula Mountain area, 60 km to the southeast of the Onts-Obo Mountain (Fig. 1, Site 9), there is an exposure of the Tavan-Khara sandstones, mudstones, limestones, tuffaceous sandstones, and gritstones, whose beds are 200 m thick in total. The bed of calcareous sandstone of variable grain-size (3 m thick) contains the assemblage identical to that from the Onts-Obo Mountain area but also including *Strophalosia khataerdensis*. Twelve meters upsection, the limestone lens (about 2 m thick) yielded brachiopods represented mostly by tropical forms characteristic of the Dzhisu-Khongor section: *Orthotetina* cf. *rubber* (Frech), *Compressoproductus* cf. *corniformis* (Chao), *Enteletes* sp., and *Prorichthofenia* cf. *ussurica* Lich. However, the assemblage lacks species widespread in the upper beds of the Dzhisu-Khongor section (Grabau, 1931).

A few years earlier, Suetenko and Ruzhentsev found the brachiopods *Echinauris* cf. *jisuensis* and *Hustedia* aff. *longirostris* Lich. together with the bryozoans *Girtiporina mongolensis* Morozova, *Maychella diserta* Gorjunova et Morozova, *Stenodiscus incertus* Morozova, and *Streblazcopora confusa* Morozova near the Khoyar-Khuduk Well of the same area (Fig. 1, Site 8). Morozova and Goryunova, who identified the bryozoans, consider them to be characteristic of the Osakhtinskaya Formation of the Khabarovsk region and the Chandalaz Horizon of the southern Primor'e region (Table 2).

Age. In the Dzhisu-Khongor sections (Dinq Yunjie *et al.*, 1985), the zonal index species was found in the Yihewusu Formation. In addition to the dominant brachiopod genera *Enteletes, Streptorhunchus, Martinia, Dielasma, Hemiptychina* and others, the assemblage of the formation contains abundant aberrant fusulinids and Midian corals. In the southern Primor'e region, the index species is known from the *Parafusulina stricta* Zone and also from the lower part of the *Metadoliolina lepida* Zone of the Chandalaz Horizon. *Stenoscisma purdoniformis* is recorded in the same units. *Anidanthus ussuricus, Horridonia morrisi*, and *Paramarginifera peregrina* occur in the Zhesi and Yihewusu formations of northern China and in the *Parafusulina stricta* Zone of the southern Primor'e region. *Neospirifer yihewusuensis* was described from the Yihewusu Formation of the Dzhisu-Khongor area. *Strophalosia* species similar to the new species *Strophalosia haraerdensis* are widespread in the upper Kazanian–lower Tatarian deposits of the Kolyma–Omolon massif and Verkhoyansk region. Some representatives of the genus *Windhamia* are recorded in Permian deposits of Australia, New Zealand, and Canada; they are also characteristic of the lower Tatarian deposits of the Verkhoyansk region and Kolyma–Omolon massif.

The facts presented above suggest the early Tatarian (lower Midian) age of the zone.

Two Late Permian basins of Mongolia were differently localized in the available mobilistic reconstructions. In "Atlas of Lithological–Paleogeographical Maps of the World" (Ronov *et al.*, 1984) based on palinspatic maps by Scotese *et al.* (1979), the basins are shown as connected with a strait crossing eastern Mongolia, but as isolated from the West Arctic basins. This interpretation was at that time in agreement with available data on the taxonomic similarity between faunas of two Mongolian basins and on the inferred migration paths of benthic invertebrates. Further studies of Permian deposits in Mongolia essentially changed the former ideas on the composition of fossil assemblages and biogeographic links. The data obtained did not confirm the hypothesis that a strait connected the Mongolian basins in the Permian time (Bobrov and Kulikov, 1968). In contrast, a close similarity was found to be characteristic of the brachiopod faunas from southern

Mongolia and the western Arctic province distinguished by Ustritskii (1971).

According to recent reconstructions (Scotese and McKerrow, 1990; Shi and Archbold, 1993; Grunt, 1995), the Siberian craton was a homeland only for the Boreal Khangai–Khentei basin of Mongolia formerly located approximately at 60°N, whereas the southern Mongolian basin was situated within the northern China (Catasian) block. The separation of the Kazakhstan block from the Siberian plate afforded the benthic fauna migration from the western Arctic into the Ussuri province.

According to the reconstruction by Shi and Archbold (1993), the Kolyma microcontinent was separated from the Siberian craton and located to the east of the Mongolian–Transbaikal province (almost at the same latitude). This reconstruction accounts for similarity between benthic faunas of the regions at the species level. However, the location of the southern Mongolian basin within the northern China block does not seem to be justified.

Ruzhentsev *et al.* (1992) presented the tectonic zoning of southeastern and inner Mongolia and suggested that the dividing line between the Siberian and Catasian plates was of a sublatitudinal direction and almost coincident with the southern boundary of Mongolia. The mentioned fauna localities in southern and inner Mongolia (the town of Yi-Ujumqin-Qi, Fig. 1), the Bei-Shan province, and Kuruk-Tag area (41°N and 92°E) were to the north of the dividing line, while the well-known Dzhisu-Khongor site of Tethyan fauna described by Grabau (1931) and other researchers (Ding Yunjie *et al.*, 1985) was to the south of it. The Kungurian faunal assemblages of southern Mongolia and the Yi-Ujumqin-Qi area consist of brachiopods, bryozoans, and bivalves characteristic of the western Arctic rather than the eastern Arctic province. The coeval Dzhisu-Khongor fauna consists of warm-water groups, such as *Monodiexodina* fusulinids[2] and scarce corals. In the period of the maximal expansion (Ufimian Age) of the most significant Permian cool-water transgression, faunas on each side of the dividing line lost fusulinids and mainly consisted of brachiopods and bryozoans similar in taxonomy to those of the *L. decimana–W. angustata* Zone described above.

The sea regression and possible paleogeographic changes of the Kazanian time disrupted biogeographic links between the northeastern Tethys and western Arctic basins. In the Ussuri province, fauna became mostly endemic.

There are no records of the early Tatarian fauna in the northern China areas bounding Mongolia in the east (the Yi-Ujumqin-Qi region) and southwest (the Bei-Shan and Kuruk-Tag regions). In southern Mongolia, discrete outcrops of the lower Tatarian deposits occur in the Solonker structural–lithologic zone and nearby. The fossil assemblages are still dominated here by brachiopods and bryozoans. Bivalves and corals occur sporadically. Brachiopods are represented by abundant *Strophalosia* and *Marginifera* forms, as well as by single representatives of typical Tethyan genera *Echinauris, Leptodus, Orthotetina*, and *Richthofenia*.

The coeval Dzhisu-Khongor assemblages are of a much more thermophilic type. In addition to the brachiopod genera listed above, they include *Enteletes, Geuerella, Meekella*, and *Uncenunellina*. Corals are abundant and diverse. The lower Tatarian (lower Midian of the Tethyan scale) deposits are characterized by aberrant fusulinids unknown in the southern Mongolia sections. The late Tatarian (late Midian) fauna is not recorded in Mongolia. Marine deposits bearing Tethyan assemblages of the late Midian age occur in the Dzhisu-Khongor area and the southern Primor'e region.

Significant differences between the early Tatarian faunal assemblages of southern Mongolia, the Kuruk-Tag, Bei-Shan, and Yi-Ujumqin-Qi areas (i.e., the faunal localities to the north of the dividing line between the Siberian and Catasian plates) and those of the Dzhisu-Khongor and southern Primor'e regions can be explained, in my opinion, by the different latitudinal positions of the corresponding basins. This explanation is supported by paleomagnetic records described by Ruzhentsev *et al.* (1992) and indicating that the Lower Permian deposits to the north of the Solonker–Solonbo–Linxi suture originally accumulated at 50°N and those to the south of the suture, at 10°N.

The quoted publication suggested the existence of microcontinents to the south of the Siberian continent. According to the palinspathic reconstructions, the Boreal Khantai–Khentei basin of northern Mongolia was located in the margin of the Siberian continent at about 60°N. It seems very probable that at 40°–50°N there existed the southern Mongolian microcontinent, whose shelf areas were inhabited by fauna less thermophilic than that of the Dzhisu-Khongor basin probably situated within the Sino-Korean plate.

According to paleomagnetic records, the sublatitudinal elongation of the Sino-Korean plate shown in the atlas by Ronov *et al.* (1984) appears to be more likely than the submeridional one suggested in the recent reconstructions (Scotese and McKerrow, 1990; and others). This is evident, for instance, from data by Likharev, Ustritskii, and Kotlyar, who studied the Dzhisu-Khongor fauna and considered it as similar to the Chandalaz fauna of the southern Primor'e region.

In conclusion, it should be noted that the suggested reference section is used to coordinate the discrete fragmental occurrences of the Late Permian fauna in different structural–lithologic zones with the strict succession of beds of this continuous section. Taxa from these

[2] E.Ya. Leven suggested the Bolorian (Kungurian) age of the assemblage after examining in 1996 the photographs of the *Monodiexodina* thin sections from the Dzhisu-Khongor area (Ding Yunjie *et al.*, 1985)

Table 2. Upper Permian Correlations

Standard Stratigraphic Scale			Mongolia		Kolyma–Omolon massif		North China	Southern Primor'e		Tethyan scale
System	Series	Stage	Horizon	Zone	Horizon	Local zone	Formation	Horizon	Local zone	Stage
Permian	Upper	Tatarian			Khivach	*Stepanoviella paracurvata*	Yihewusu	Chandalaz	*Neomisellina lepida–Lepidolina kumaensis*	Midian
			Solonker	*Echinauris jisuensis*	Gizhiga	*Cancrinelloides curvatus*			*Parafusulina stricta*	
						Cancrinelloides obrutchewi			*Monodiexodina sutschanika*	
		Kazanian	Tsagan-Ula	*Alispiriferella lita – Kaninospirifer adressum*	Omolon	*Magadania bajkurica*	Zhesi	Vladivostok		Murgabian
						Terrakea korkodonensis				
						Terrakea borealis				
		Ufimian		*Liosotella decimana – Waagenoconcha angustata*		*Omolonia snjatkovi*				Kubergandian
						Mongolosia russiensis				
						Kolymaella ogonerensis				

occurrences considerably completed the stratotype assemblages.

The brachiopod zones distinguished are based on significant changes in assemblages, which were caused by different abiotic events. For instance, the *L. decimna–W. angustata* Zone corresponds to deposits accumulated during the maximum of the cool-water transgression of the Kungurian–Kazanian time. The brachiopod assemblage of the zone includes many boreal genera and species characteristic of coeval assemblages from the western Arctic province. As the sea retreated and the cycle changed (the *A. lita–W. angustata* and *E. jisuensis* zones), the endemic species increased in abundance in the Ussuri province. According to Miklukho-Maklai (1963), this province comprises southern Mongolia, northeastern China, the Far East, and Japan.

The article presents the first description of the latest Permian brachiopod assemblage from the region under consideration, i.e., that of the *E. jisuensis* Zone. It is characteristic that the assemblage includes species first described from coeval deposits of the Boreal basin of northeastern Mongolia. This may suggest that close biotic connections existed between the basins in the Late Permian time.

In addition to transgression and regression, there were other abiotic factors that essentially affected biota of the basins. Similarly significant were changes in the water temperature, sedimentary facies, shelf areas, directions of fauna migration, and others. For instance, thermal parameters of the past basins and paleolatitude determinations were used to substantiate the existence of one or several microcontinents. Shelf zones of the latter were inhabited in the Permian time by the peculiar "moderately warm-water" fauna of the mixed Boreal–Tethyan type, which was characteristic of southeastern Mongolia, northeastern China, and some Primor'e areas to the north of the Solonbo–Linxi suture.

Reviewers L.A. Nevesskaya and A.Yu. Rozanov

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Stages of the Early Triassic Ammonoid Evolution in the Eastern Boreal Province

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Received January 13, 1997; in final form, November 10, 1997

Abstract—The history of the evolution of the Early Triassic ammonoid communities in the eastern Boreal province is discussed with due regard for the peculiarities of their biogeographic differentiation and phylogenetic relations. The evolution is shown to include two rhythms of fauna development in the Induan and Olenekian stages, four stages, and 11 phases corresponding to 22 zones and subzones of the biostratigraphic zonation of the Lower Triassic in southeastern Asia.

Key words: *Early Triassic, ammonoids, rhythm, stage, substage, phase, zone.*

The Early Triassic ammonoid communities of the eastern Boreal province included immigrants and species representing several endemic phylogenetic lines. Some immigrants were ancestral for these endemic lines. Not all phylogenetically related species were truly endemic. Some of them were emigrants, which migrated into other Boreal basins and the Tethys from a speciation center located in the eastern part of the Boreal province. Since the composition of ammonoid communities in the eastern Boreal basin was mixed, the history of their development is considered with due allowance for the peculiatities of geographic differentiation and for the phylogenesis of all ammonoids of the basin. The history of the development of ammonoids is accepted as a basis for biostratigraphic zonation of the Lower Triassic in northeastern Asia (Dagys and Ermakova, 1993; Dagys, 1994).

In biogeographic terms of Enay (1980), the Early Triassic of the eastern Boreal province reveals two classical rhythms of ammonoid fanna development, which correspond to two Lower Triassic stages, i.e., the Induan and Olenekian. Each rhythm consisted of two periods spanning the substages. As a rule, the first period was a time of the appearance and flourishing of cosmopolitan ammonoids, whereas the second was characteristic of an endemic community. The substages are divided into several phases of development of one or several genera and spanning one or several zones of the Lower Triassic biostratigraphic zonation. The phases may include smaller subdivisions corresponding to the evolution time of one or several species.

The evolution of the Induan ammonoids had two substages (Fig. 1). In the early Induan time of a global transgression, the ammonoids were diversified little, and the common genera dominated in basins of different paleolatitudes (Dagys and Ermakova, 1996). In the eastern part of the Boreal province, two invasions of genetically incoherent groups of ammonoids probably appeared at the first substage of the first rhythm. The first invasion was that of the *Otoceras* genus, if arctic Canada is assumed to be a center of its dispersal. However, neither arctic Canada, nor eastern Verkhoyansk regions provide data for tracing a connection of *Otoceras concavum* with its ancestral forms. Therefore, the problem of the *Otoceras* dispersal center cannot be definitely solved. If this center was in the eastern Verkhoyansk region, then the migration of ammonoids to the eastern part of the Boreal province could be triggered at the beginning of the period, as also the invasion that followed and caused the appearance of representatives of the *Tompophiceras* genus and all ophiceratids genetically associated with it. The radiation center of *Tompophiceras* was in the western part of the Boreal province. Other Boreal regions do not show such a distinct differentiation between the *Otoceras* and the *Ophiceras* faunas. In eastern Greenland (Spath, 1930; 1935), arctic Canada (Tozer, 1994) and Spitsbergen (Korchinskaya, 1986), the *Otoceras*, *Tompophiceras*, and ophiceratids coexisted within the *boreale* Zone.

The majority of the early Induan genera from northeastern Asia are immigrants, which is probably the cause of a limited applicability of phylogenetic methods to this period. Only short phyletic lines of species from the *Otoceras*, *Tompophiceras* and *Wordieoceras* genera can be detected. Nonetheless, two early Induan phases are identified in the ammonoid evolution, and the subdivision of this substage is based on the phylogenetic lines of the *Otoceras*, *Tompophiceras*, and *Wordieoceras* genera. The first phase of the early substage in development of the Early Triassic ammonoids correlates with the *concavum* and *boreale* zones and is characterized by the appearance of *O. concavum* and

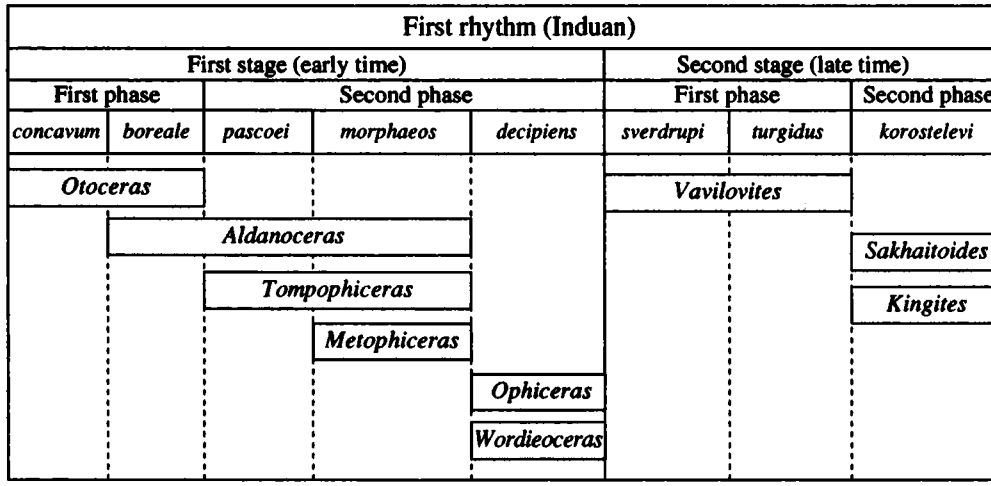

Fig. 1. The evolutionary stages of the Induan ammonoids.

O. boreale, the latter representing a result of evolutionary transformations of the former. Moreover, at the end of that substage, the *Aldanoceras* genus appears as the only taxon of the Triassic xenodiscids. In the western part of the Boreal province, *Tompophiceras*, *Ophiceras*, and *Vishnuites* forms coexisted with *O. boreale* during the middle of the first phase.

The second phase of the early substage spans the *pascoei, morphaeos,* and *decipiens* zones. Its beginning in the eastern part of the Boreal province coincides with the origin of two western Boreal species of the *Tompophiceras* genus: *T. pascoei* and *T. gracile*. Their evolution gave rise to the endemic species *T. morphaeos* and *T. bychkovi*. In the middle and at the end of this phase, there appeared two immigrant genera associated genetically with the genus *Tompophiceras*. In the eastern Boreal area, each of them is represented by one species only, namely by *Metophiceros subdemissum* and *Ophiceras transitorium*. The former species initiated the philogenetic series that developed in the late Induan and early Olenekian. The second phase terminated with the appearance of yet another western Boreal *Wordieoceras* genus represented by species *W. decipiens*, the evolutionary transformations of which resulted in the appearance of *W. tompoense*.

In the second substage of the first rhythm, apparently in connection with the decrease of the eustatic level of the World Ocean, the geographic differentiation of marine invertebrate faunas, ammonoids included, suddenly increased. At that time, basins located at different latitudes were in fact populated by various types of faunas. There is a certain indication of isolation of the eastern and western parts of the Boreal province. In in the east of the Boreal area, two late Induan phases become evident in the ammonoid evolution. The first of them corresponded to the time of the *Vavilovites* genus development. The biozonation of the lower part of the upper Induan Substage is based on the evolution of this genus demonstrating a gradual change of taxa from the relatively thin forms with the narrow ventral side to the swelled species (particularly, near the umbilical part) with the wide ventral side. The first phase corresponds to the *sverdrupi* and *turgidus* zones (*subtriangularis* and *umbonatus* subzones of the latter) and is characterized by the absolute predominance of the *Vavilovites* genus. In the western part of the Boreal province, the *Proptychites* genus was prolific along with the early *V. sverdrupi* species, and the immigrant genus *Prionolobus* arrived from Tethys. The *Vavilovites*, forms of a later period which are widespred in northeastern Asia, are unknown in western Boreal regions.

The second phase of the late Induan substage corresponds to the *korostelevi* Zone and is characterized by the appearance of the endemic *Sakhaitoides* genus apparently associated in genesis with *Metophiceras* forms and represented by two species; the *korostelevi* species that appeared simultaneously are tentatively referred to the *Kingites* genus. The origin center of the *Kingites* genus is unknown so far. In the western part of the Boreal province, its representatives coexisted with the *Vavilovites* and newly appearing forms of the *Koninckites* and *Flemingites* genera. Moreover, Tozer (1994) reported from Arctic Canada about the coeval *Xenodiscoides calnanani* and *Wyomingites scapulatus* species, whose systematic position is debatable.

The second rhythm of ammonoid evolution, as the first one, included two evolutionary phases of both the Boreal and the low-latitude ammonoids (Zakharov, 1975). Each of these phases spans the time interval corresponding to a substage (Fig. 2). In the early Olenekian, an extensive invasion of the Tethyan taxa into the Boreal basins was recorded, and cosmopolitan genera dominated in the ammonoid communities, which is in good agreement with data on the global transgression that occurred at that time. In this case, as in the first

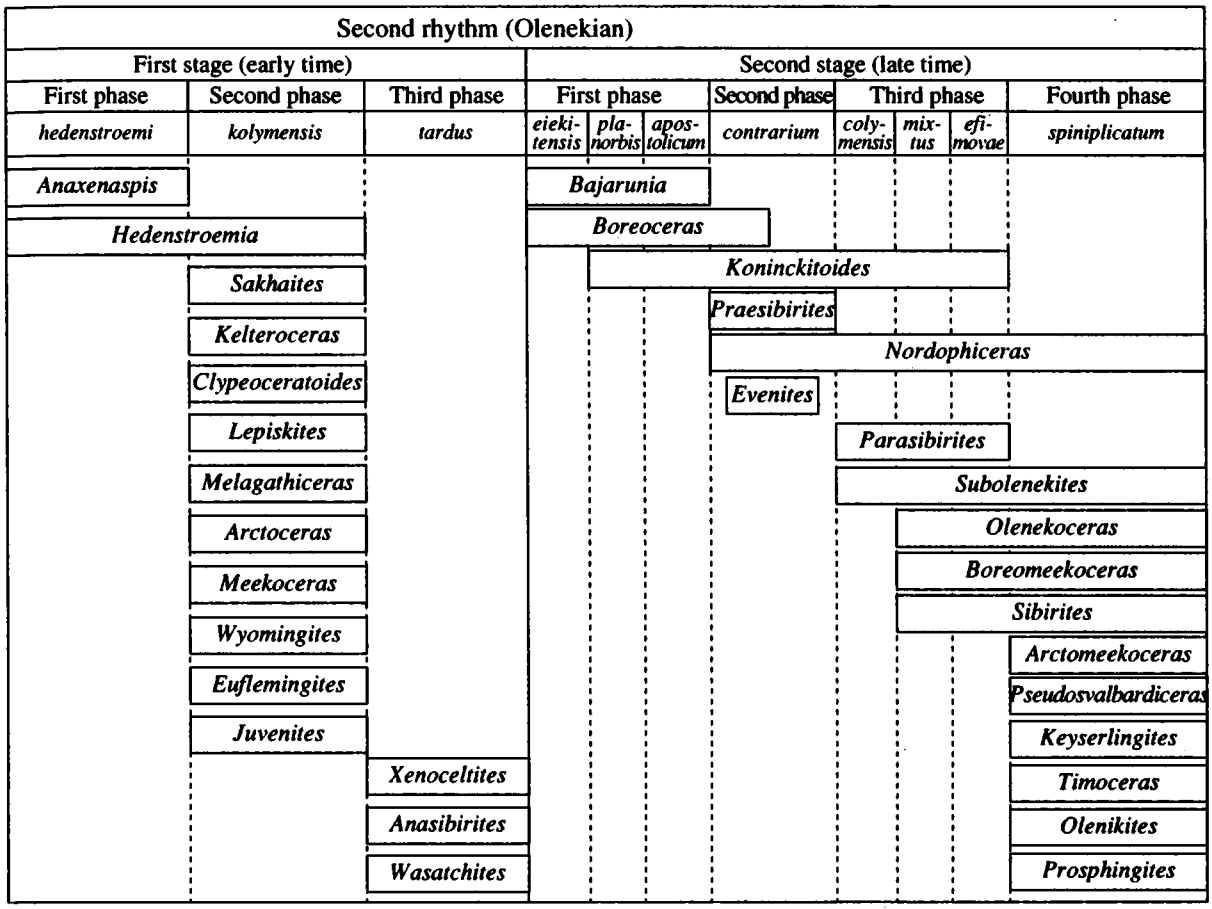

Fig. 2. The evolutionary stages of the Olenekian ammonoids.

rhythm, there could have been the emigration of ammonoids at the beginning of the period, if we assume the eastern Boreal province as a center of the hedenstroemiid dispersal. But even if the Himalayas were that center, the appearance of hedenstroemiids in the western Boreal area was associated with their migration from the eastern part. Concurrently with the cosmopolitan genera, individual endemic phylogenetic lines likewise developed in the eastern part of the Boreal province. Using the historical development of the eastern Boreal ammonoid communities as a basis, the first stage of the second rhythm can be divided into three phases.

The beginning of this period was associated with the appearance of two cosmopolitan *Hedenstroemia* and *Anaxenaspis* genera. In all probability, the eastern part of the Boreal province was actually the center of origin and dispersal of these genera, if my supposition about their origin is correct. The most likely ancestor of hedenstroemiids is the late Induan *Kingites* genus, whereas the roots of the *Anaxenaspis* genus must be among the evolute forms that departed from the *Vavilovites* genus at the beginning of the late Induan. The first phase of the first stage of the second rhythm corresponds to the *hedendstroemi* Zone and is characterized by the domination of *Hedenstroemia hedenstroemi* and genetically associated *H. tscherskii*. The evolution of hedenstroemiids progressed toward the complication of their lobe line. The appearance of *Anaxenaspis olenekensis* was concurrent with that of the hedenstroemiids. The phase may also mark the first occurrence level of *Sakhaites* taxa known only as fragments recovered at a single site at the Vendyavkiri Brook, the left-hand tributary of the Buur River. This supposition, however, is premature without supporting data from other sections (Dagys and Ermakova, 1990). In the eastern part of the Boreal province, hedenstroemiids were recovered from two sites in arctic Canada (Dagys and Tozer, 1989; Tozer, 1994); in all other regions, a paleontological hiatus is recorded as corresponding to the evolution time of ammonoids at the phase in question.

The second phase of the early Olenekian substage corresponds to the *kolymensis* Zone and has been distinguished by an extensive invasion of cosmopolitan genera such as *Arctoceras, Meekoceras, Wyomingites, Euflemingites, Melagathiceras*, and *Juvenites*. The phylogenetic lines of concurrent endemic genera *Lepiskites* and *Clypeoceratoides*, which were genetically

related to the genus *Anaxenaspis*, likewise appeared; their evolution was characterized by the complication of the lobe line and by a greater degree of the shell convolution. The *Sakhaites* and *Kelteroceras* genera were genetically related to the late Induan *Sakhaitoides* and evolved toward the simplification of their morphology, mainly in the lobe line structure. The *Hedenstroemia* genus terminated its evolution and was represented in the phase under consideration only by a single species described in the open nomenclature. It should be noted that the endemic phylogenetic lines of the eastern Boreal ammonoids, the origin of which goes as far back as the Induan, completed their evolution in this phase. The phase is distinguished by the greatest taxonomic diversity of the Early Triassic ammonoids in the eastern part of the Boreal province. In the western part of the Boreal basin, the second phase corresponds to the *romunderi* Zone as diverse in the taxonomic aspect as that in the eastern Boreal province. Apart from genera in common (*Meekoceras, Arctoceras, Euflemingites, Juvenites, Melagathiceras*), the western basin was populated by *Kashmirites, Oxyussuria*, and *Paranannites* forms, which were probably immigrants from the Tethys.

The third phase of the early Olenekian Substage corresponds to the *tardus* Zone characterized exclusively by cosmopolitan *Wasatchites*, *Anasibirites*, and *Xenoceltites* genera. In other words, the invasion of ammonoids from the Tethys and western Boreal province continued at this phase. The *Wasatchites* and *Anasibirites* forms were not abundant in the eastern part of the Boreal area. The *Xenoceltites* genus of two taxa (*X. subevolutus* and *X. matheri*) was obviously dominant, and its evolution toward a more complicated lobe line and changes in ornamentation produced two endemic branches that extensively developed in the second substage of the rhythm. Ammonoids of the considered third phase were not considerably diverse. In the western Boreal basin, the third phase corresponds to the *tardus* zone characterized by a similar ammonoid community, though of a larger taxonomic diversity.

The late Olenekian substage of the second rhythm of ammonid evolution corresponded to the time of their excessive endemism and greater geographic differentiation; this time probably marked an essential lowering of the World Ocean level. Sea basins in the lower and higher latitudes were populated by different types of faunas, and there were only certain short periods when individual boreal genera penetrated into the boundary Tethyan basins and, vice versa, when the Tethyan forms appeared in the boreal ammonoid communities. There were four phases of the ammonoid evolution in the eastern part of the Boreal province during the second Olenekian Substage. The first phase corresponds to the *eikitensis, planorbis* and *apostolicum* zones. The beginning of the phase is marked by the simultaneous appearance of two endemic genera *Bajarunia* and *Boreoceras* phylogenetically associated with the different species of the *Xenoceltites* genus. The evolution of the *Bajarunia* genus, represented by six species (*B. eiekitensis, B. taimyrensis, B. alexandri, B. eumphala, B. alexeevae*, and *B.* sp.), is confined exclusively to the first phase. The evolutionary transformations of the genus are recorded in insignificant complications of the lobe line. Three *Boreoceros* species (*B. planorbis, B. apostolicum,* and *B. demokidovi*) do not occur beyond the three aforementioned zones. In the middle of the first phase, another endemic genus *Koninckitoides* appeared as the terminal one in the phylogenetic line *Anaxenaspis–Lepiskites–Clypeoceratoides*. The appearance of this genus was associated with the simplification of the taxa organization. In the western Boreal regions, ammonoids of the first phase are unknown, and a paleontological hiatus spanning three zones is inferred for this province.

The second phase of the substage corresponds to the *contrarium* Zone divided into the *lenaense, tuberculatus*, and *egorovi* subzones. Its beginning coincides with the appearance of the *Nordophiceras* genus departed from the *Bajarunia* line because of evolutionary transformations that resulted in the disappearance of peculiar sculptural elements of the latter. In the first half of the second phase, the *Boreoceras* genus (*B. lenaense* and *B. mirabile*) terminated its evolution; the evolutionary changes of the genus toward the enhancement of sculptural elements resulted in the appearance of another endemic genus *Praesibirites*. The evolution of this genus, represented by two species (*P. tuberculatus* and *P. egorovi*), was confined entirely to the second phase, when it coexisted with the *Koninckitoides* genus. In the middle of this phase, there appeared the *Evenites* genus, an immigrant taxon from the Tethys, whose existence time is limited by the zonal *tuberculatus* event. In the western part of the Boreal province, the ammonoides of the second phase, like those of the first, are unknown.

The appearance of *Subolenekites pilaticus*, an immigrant species from western regions of the Boreal province, and the advent of the *Parasiberites* genus, as a result of evolutionary transformations of *Praesibirites*, marked the beginning of the third phase of the late Olenekian. It corresponds to the *kolymensis, mixtus*, and *efimovae* zones. At that phase, four species of the *Subolenekites* genus originated. Two of them (*S. pilaticus* and *S.? shevyrevi*) were ancestral for the *Olenekoceras* and *Boreomeekoceras* genera. The evolution of the *Parasibirites* genus (*P. grambergi, P. kolymensis, P. subpretiosus, P. mixtus, P. efimovae*) is confined exclusively to the third phase. All the changes in the genus shells concern only different sculptural elements; the duration of the sculptural transformations; and, insignificantly, the shell shape. In the middle of this phase, the *Sibirites* genus departed from the *Parasibirites* line and evolved separately, displaying a simplification of sculptural elements and the lobe line. Two species (*S. elegans* and *S. pretiosus*) appeared at different moments of the third phase, when *Nordophiceras* forms continued their evolution. At the beginning of the

phase, two new-comers (*Nordophiceras menense* and *N. kasakovi*) appeared amid the latter and existed until the end of the phase. The *Koninckitoides* genus is terminal in the phylogenetic branch, whose origin goes back to the late Induan and is completed within the phase. This phase of the ammonoid evolution is also traceable in the western part of the Boreal province (arctic Canada), where only one species *Subolenekites pilaticus* is reliably established and initially assigned by Tozer (1967) to the *Olenikites* genus. Later, Tozer (1994) described the new species *Olenikites subtilis* that belongs in all probability again to the *Subolenekites* genus together with *Neomeekoceras scalariforme* practically identical to *Nordophiceras menense* and, apparently, occupying the same position in the ammonoid systematics.

The fourth phase of the late Olenekian corresponds to the *spiniplicatus* Zone. In the degree of taxonomic diversity of ammonoids, it differs only slightly from the second phase of the early Olenekian substage. The beginning of the phase coincides with the appearance of as many as five new endemic genera: *Olenikites, Timoceras, Pseudosvalbardiceras, Keyserlingites*, and *Arctomeekoceras*. At this time, the *Nordophiceras* genus continued its evolution by producing three new species (*N. karpinskii, N. popovi, N. schmidti*). The *N. ordophiceras* genus evolution, which resulted in the flattening of ventral shell sides, produced a new endemic genus *Pseudosvalbardiceras*, the taxon important for further development of ammonoids in the Middle Triassic. The fourth phase of the late Olenekian was a time of an extreme diversification in the *Olenekoceras* line, when *O. middendorffi* became extinct and two new species of the genus (*O. shrenki, O. nikitini*) appeared. The evolution of the genus increased the shell size and resulted in the origin of more prominent sculptural elements and a more intricate lobe line. As a consequence of further development of lobe lines and new sculptural elements in this direction, the monotype *Keyserlingites* genus departed from the *Olenekoceras* line and spread throughout the entire Boreal area. Two other monotype genera (*Timoceras* and *Olenikites*) appeared as a result of evolutionary transformations of *Subolenekites altus* that terminated its evolution at this stage. The *Arctomeekoceras* genus appeared at the beginning of the stage. The *Sibirites* and *Boreomeekoceras* genera terminated their evolution at this stratigraphic level, and the *Proshingites* genus, an immigrant from the Tethys, appeared. In the western part of the Boreal province, the fourth phase coincides with an extensive invasion of Tethyan ammonoids. Of a large ammonoid assemblage of *Preflorianites, Proshingites, Isculitoides, Popovites, Zenoites, Monacanthites, Metadagnoceras, Svalbardiceras, Keyserlingites, Procarnites*, and *Leiophyllites* taxa, only two species (*Keyserlingites subrobustus* and *Proshingites czekanowskii*) are known to be inherent to both parts of the Borael province together with a few species attributed by Tozer (1994) to the *Olenikites* genus.

The analysis given above shows that the biostratigraphic zonation of ammonoids from northeastern Asia corresponds, in general features, to the rhythms, stages and phases of their evolution in the eastern part of the Boreal province. The established rhythms correspond to stratigraphic stages; the lesser periods, to substages; and phases, to one or several zones. A similar stage sequence in the ammonoid evolution is also established for the western part of the Boreal province. Consequently, an assumption that the rhythms and lesser evolutionary periods detectable in the Boreal province may correspond to the global phases of the ammonoid evolution cannot be ruled out.

Reviewers A.S. Alekseev and A.A. Shevyrev

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Biostratigraphic Subdivision of Mesozoic Deposits Penetrated by the Tyumen Superdeep Borehole

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Received March 18, 1997

Abstract—The Tyumen superdeep borehole (TSB-6) drilled in north-western Siberia was the first that penetrated through Mesozoic deposits and probably entered the uppermost Paleozoic sediments. In the course of drilling, a clear succession of stratigraphic units, established formerly in separate Mesozoic sections of western Siberia, was revealed. The complete section of Triassic volcano–sedimentary deposits underlying the Jurassic sedimentary strata with no visible unconformity was recovered for the first time. Paleontological remains, which had recently been obtained from the borehole section, was used to substantiate biostratigraphic subdivision of the Mesozoic deposits penetrated by TSB-6, and to constrain the age of stratigraphic units and their stratigraphic ranges. Palynospectra characteristic of the Middle–Late Triassic, Jurassic, and earliest Cretaceous time allowed a wide chronostratigraphic correlation to be carried out within and beyond the Urengoi region. On the basis of macrofossil plant assemblages, the correlation of the TSB section with the reference borehole sections of the Urengoi area is suggested. Microfauna assemblages from coastal-marine interlayers occurring inside the Mesozoic continental sequence suggest the peculiar dynamic environments during the whole period of the Mesozoic sedimentation.

Key words: *superdeep borehole, Urengoi area, biostratigraphy, Mesozoic, palynospectra, microfauna, plant macrofossils, correlation, stratigraphic units, formations.*

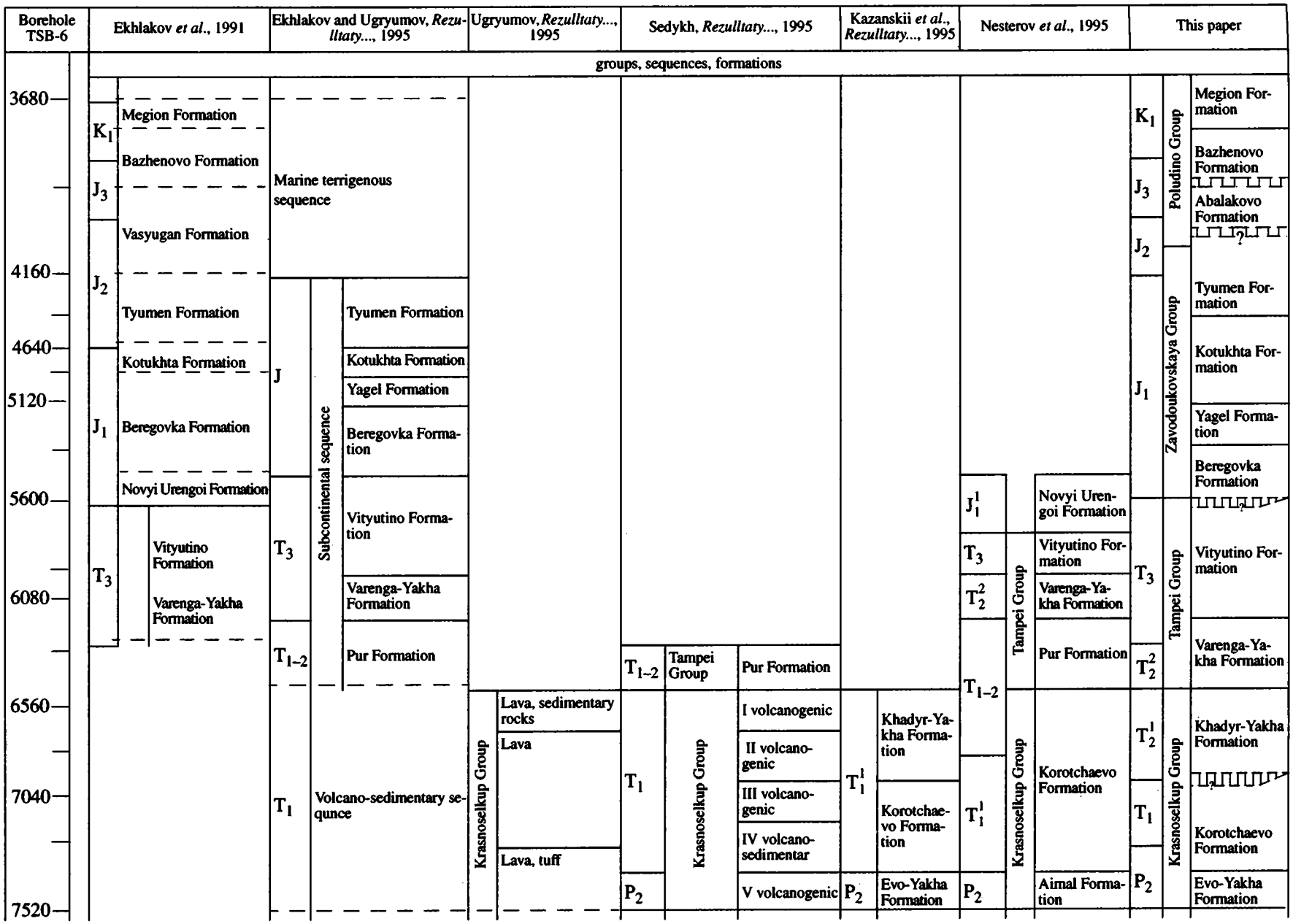
The Tyumen superdeep borehole (TSB) is located in the north of western Siberia, in the Urengoi oil-bearing area of the Nadum–Pur oil-and-gas province. The drilling was commenced in 1987 to reach in 1994 the depth of 7502 m, the maximum one for sedimentary basins. Down to the depth of 3698 m, the drilling proceeded without core sampling. The depth interval of 3698–4100 m was sampled completely, and deeper, the core recovery was 80%. The drilling resulted in obtaining unique material that allows a reliable verification of many questions, including the litho- and biostratigraphic characteristics of the Mesozoic deposits in the Urengoi area perspective for oil-and-gas deposits.

The TSB section interval of 3698–7502 m is represented by continental, marine, and coastal sediments of the Cretaceous, Jurassic, and Triassic age, as well as, probably, by Upper–Middle Triassic and Permian volcanogenic deposits. Initial data on the stratigraphic subdivision of the deposits in the depth interval of 3698–6194 m were published by Ekhlakov *et al.* (1991). According to the stratigraphic scheme, adopted for the Urengoi area at the Interdepartmental Stratigraphic Conference (ISC) held in Tyumen in 1990 (see table), the penetrated sequence is subdived from the base upward into the following units: the Tampei Group with the Varenga-Yakha and Vityutino formations (the latest Middle–Late Triassic); the Zavodoukovskaya Group with the Novyi Urengoi, Beregovka[1], Kotukhta (Early Triassic), Tyumen (Middle Jurassic) formations; the Poludino Group with the Abalakovo (Middle and the beginning of the Late Jurassic), Bazhenovo (Late Jurassic, earliest Cretaceous), and Megion (the Achimovka Member) formations of the Early Cretaceous (Berriasian). These stratigraphic units were dated on the basis of preliminary data on spore–pollen assemblages studied by S.I. Purtova.

In February, 1995, the Roskomnedra, NPP "Nedra," and KamNIIKIGS initiated a meeting devoted to the TSB drilling results, which was held in Perm. Problems of drilling equipment and technology, the deep structure, and oil-and-gas potential were the subject of wide discussion at the meeting. The stratigraphic subdivision of the penetrated sequence, especially, biostratigraphy, age substantiation, and correlation of deposits within the region, were almost missing from discussion. Only abstracts presented by Ekhlakov, Ugryumov, Sedykh, and Kazanskii (*Rezul'taty...*, 1995) were devoted to subdivision (table) of the lower volcano-sedimentary part of the sequence into lithostratigraphic units on the basis

[1] The names of the formations were used loosely by the cited authors, and, according to the scheme adopted at the ISC in Tyumen in 1990, the lower unit of the Jurassic sequence is the Beregovka Formation (the name *Novyi Yrengoi* appeared to be preoccupied), and the unit overlying it is the Yagel Formation.

Subdivision of the Mesozoic deposits penetrated by TSB-6



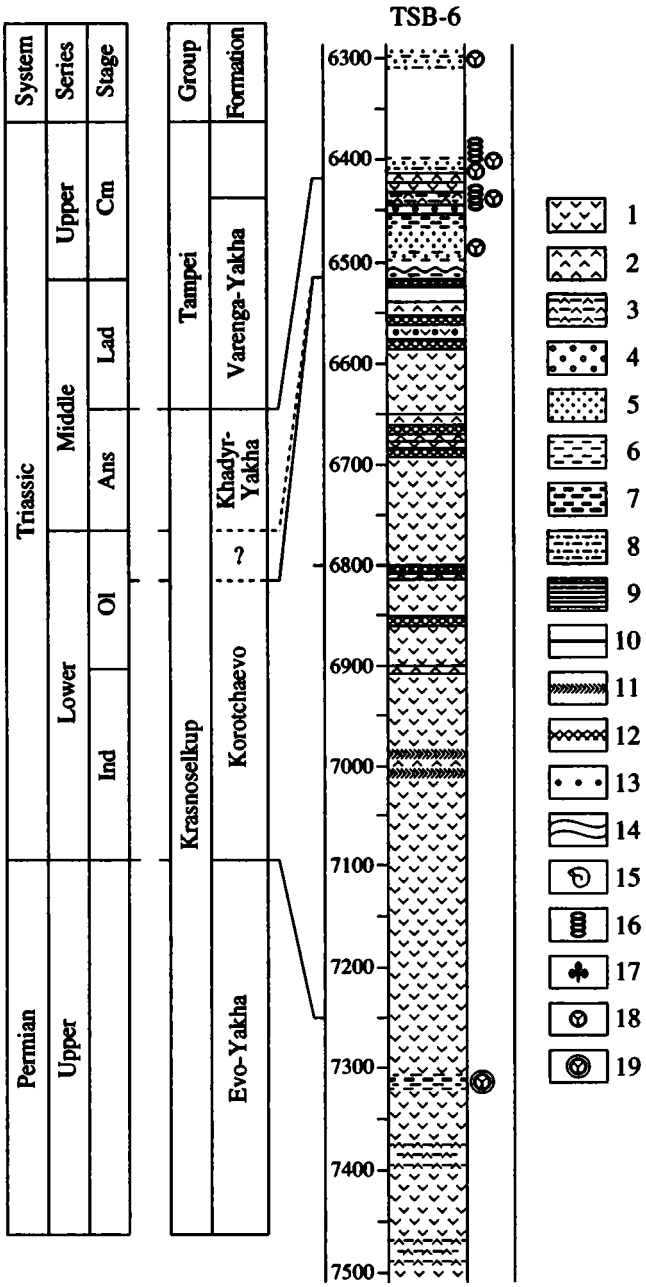


Fig. 1. Subdivisions of Permian–Triassic deposits penetrated by TSB-6:
(1) extrusive rocks; (2) tuff; (3) tuffite; (4) conglomerate; (5) sandstone; (6) mudstone; (7) coaly mudstone; (8) siltstone; (9) clay; (10) coal; (11) quartz–epidote rocks; (12) weathering crust; (13) laterite; (14) void; (15) ammonites; (16) foraminifers; (17) plant macroflora; (18) microphytofossils; (19) Permian palynoassemblage (after Purtova).

of palynological data (communication by Purtova in *Rezul'taty* ..., 1995).

The paper by Nesterov *et al.* (1995), which appeared a bit later, presented a subdivision scheme, where the lower part of the TSB sequence was divided into two units (table): the Krasnoselkup Group with the Aimal and Korotchaevo formaions, and the Tampei Group with the Pur, Varenga-Yakha, and Vityutino formations. Age of these stratigraphic units was again determined on the basis of palynological data. Unfortunately, the paper presents only general characteristics of pollen assemblages and their names. This hampered the use of palynological data for the biostratigraphic analysis and age determination of deposits.

In recent years, by the courtesy of A.A. Dikovskii, chief geologist of the Tyumen Prospecting Expedition on Superdeep Drilling, NPP "Nedra," we also were supplied with geological maps and cross sections; we were also given the opportunity to examine and describe the TSB core and to take paleontological samples. The investigation and subsequent analysis of the material with due regard for the data published are used as a basis for the work presented.

The lithologic column (Figs. 1–3) was compiled on the basis of core description involving the data presented by geologists from the Tyumen Prospecting Expedition. The data processing, especially, the identification of paleontological samples, appeared to be the most time-consimung procedure. The high-grade metamorphism of rocks often represented by coarse-grained sandy varieties, as well as peculiar conditions of sedimentation, affected greatly the extent of fossil preservation. This is especially true of microfossils—spores, pollen, microphytoplankton and microfauna. Plant macrofossils have suffered the destructive effect of metamorphism as well. Separate interlayers comprise only imprints of stems and leaves but no phytoleims (mummified leaves), which prevented the study of epidermal structure of Gymnospermae leaves.

PERMIAN(?), TRIASSIC

The *Krasnoselkup Group* of the TSB section comprises a volcano-sedimentary sequence more than 1000 m thick occurring at the base of Mesozoic strata (*Rezul'taty* ..., 1995). There is no consensus regarding the subdivision of the group. Nesterov *et al.* (1995) proposed to subdivide the group into two formations (table): the Aimal (the Late Permian) and Korotchaevo (the Early Triassic, Induan) formations. We consider the suggestion by Kazanskii and co-authors as more justified (*Rezul'taty* ..., 1995; Kazanskii *et al.*, 1995). They subdivide the Krasnoselkup Group into three units: the Evo-Yakha, Korotchaevo, and Khadyr-Yakha formations.

The *Evo-Yakha Formation* is basal in the group (the depth interval of 7502–7250 m, Fig. 1). It is represented by lava and tuff beds with sills, dikes, and thin intercalations of tuffites and coaly mudstone. Purtova identified the Permian assemblage of palynomorphs within the depth interval of 7317–7307 m (Kazanskii *et al.*, 1995).

The *Korotchaevo Formation* conformably overlying the Evo-Yakha Formation (Kazanskii *et al.*, 1995)

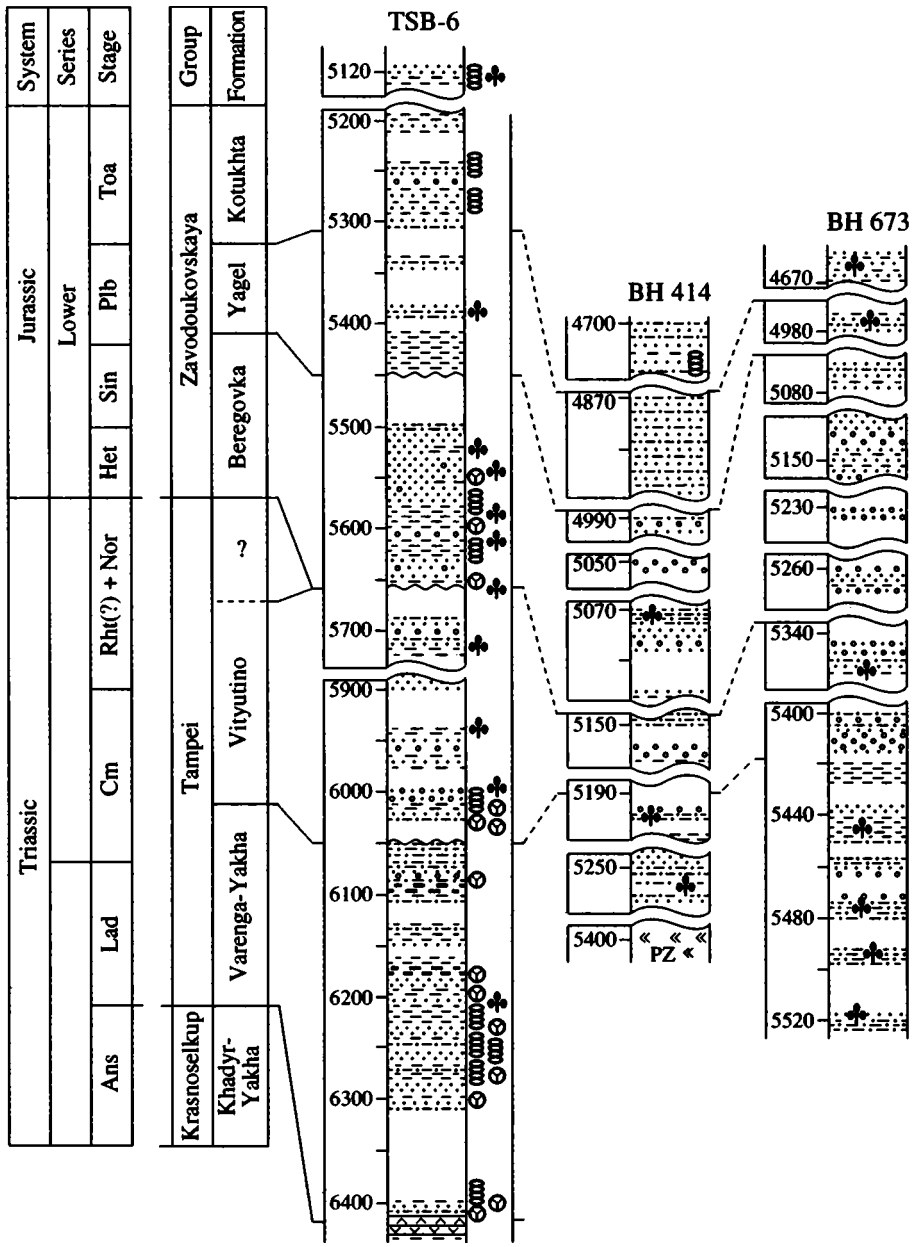


Fig. 2. Subdivision scheme of Triassic and Lower Jurassic deposits penetrated by TSB-6 and correlation with the reference borehole sections of the Urengoi area (symbols as in Fig. 1).

is distinguished within the depth interval of 7250–6520 m. The formation consists of two units. The lower subformation (depth interval of 7250–6800 m) is represented by altered extrusive rocks and tuffs with interlayers of weathering crust and quartz–epidote rocks; the upper subformation (the depth interval of 6800–6520 m) is composed of the same rocks and laterites (Kazanskii *et al.*, 1995). The formation was attributed to the Induan stage of the Lower Triassic, though, as is shown below, the lower beds of the Khadyr-Yakha Formation overlying the Korotchaevo Formation are characterized by the Middle Triassic spore–pollen assemblage (arbitrarily the Anisian). Most likely, the Korotchaevo Formation spans not only the Induan, but also certain Olenekian strata of the Lower Triassic.

The *Khadyr-Yakha Formation* is distinguished by Kazanskii *et al.*, (1995) within the depth interval of 6520–6421 m. The formation occurs discordantly over the Korotchaevo Formation. It is represented by alternating claystone and graywacke sandstone beds at the base, which are overlain by alternating tuffs, tuffites, and altered extrusive rocks. On the basis of palynological data by Purtova, Kazanskii *et al.* (1995) referred the formation to the Lower Triassic Induan Stage,

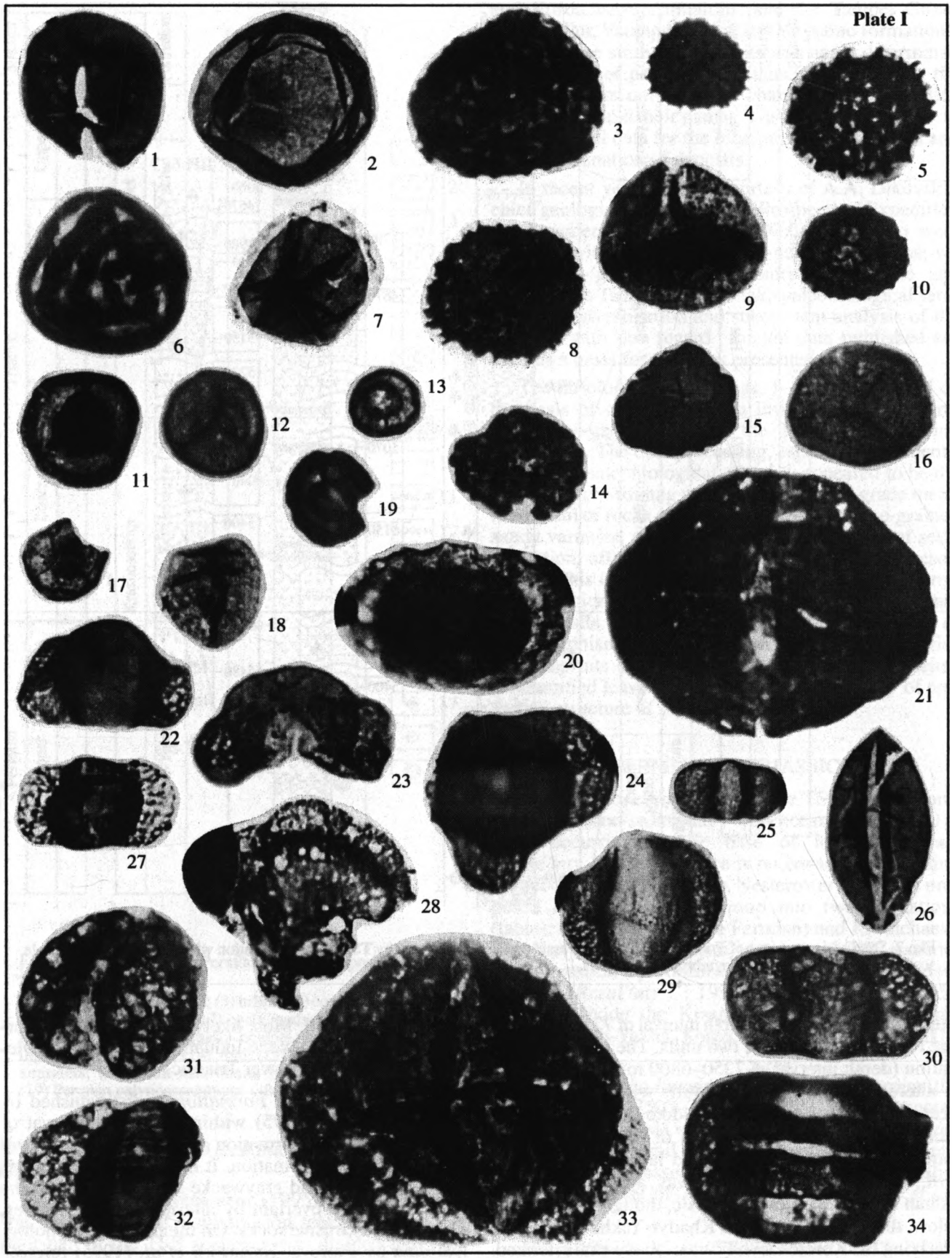


Plate I
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although palynological data obtained by Kulikova for almost the whole section of the Khadyr-Yakha Formation (Fig. 1) prove the Middle Triassic age of the formation. The prevalence of pollen of conifers over spores is characteristic of palynomorphs from this formation (Plate 1). Dominant pollen species are *Alisporites* sp. and *Latosaccus latus* Mädler; they associate with *Platysaccus reticulatus* Mädler, *Umbrosaccus* sp., *Sulcatisporites kraeuseli* Mädler, *Colpectopollenites ellipsoides* Visscher, *Falcisporites snopkovae* Visscher, *Stellapolenites thiergartii* (Mädler) Clement-Westerhof, *Klausipollenites* sp., *Striatoabietites aytugii* Vischer, and *Taeniaesporites pellucidus* (Goubin) Balme. Spores of the assemblage are represented by *Calamospora* sp., *Punctatispora* sp., cf. *Spinotriletes echinoides* Mädler, *Perotriletes minor* (Mädler) Clement-Westerhof, *Duplexisporites* sp., *Cyclotriletes* sp., *Densoisporites* sp., *Nevesisporites fossulatus* Balme, *Retusosporites mesozoicus* Klaus, *Apiculatisporites* sp., *Rugulatisporites mesozoicus* Mädler, and *Aratrisporites* sp. The stratigraphic range of many defined species is mainly confined to the latest Early–Middle Triassic, whereas species *Stellapollenites thiergartii, Perotriletes minor, Platysaccus reticulatus, and Latosaccus latus* are known in western Europe only from the Middle Triassic beds (Mädler, 1964; Visscher and Brugman, 1981). Moreover, in the taxonomic aspect, the Khadyr-Yakha palynological assemblage is similar to that from the Anokhino Formation (Assemblage V) of the Tura Group of the Anokhino depression of the eastern Urals, which was referred by Tuzhikova (1973) to the Anisian Stage of the Middle Triassic. These are grounds to assign the Khadyr-Yakha Formation to the Middle Triassic as well (arbitrarily, to the Anisian).

Fossil macrofauna of poor preservation was found in the middle part of the Khadyr-Yakha Formation. L.L. Ovchinnikova identified *Ostracoda* sp., *Hyperammodiscus* (?) sp., and rare undeterminable foraminifers within the depth interval of 6488.4–6480.2 m, as well as *Hyperammina* sp. and *Ortovertella* ? cf. *coctilis* Schleifer within the depth interval of 6457–6447 m. The remains of ostracodes and foraminifers are found mainly in volcanogenic deposits and indicate a short existence period of a shallow sea in the Urengoi region even in the Middle Triassic.

Volcano-sedimentary deposits of the Krasnoselkup Group are overlapped by the continental sedimentary sequence with rare interlayers of coastal-marine sediments. The deposits are subdivided into the Tampei (the latest Middle–Late Triassic), Zavodoukovskaya (the Early–Middle Jurassic), and the Poludino (the latest Middle–Late Jurassic—earliest Cretaceous) groups.

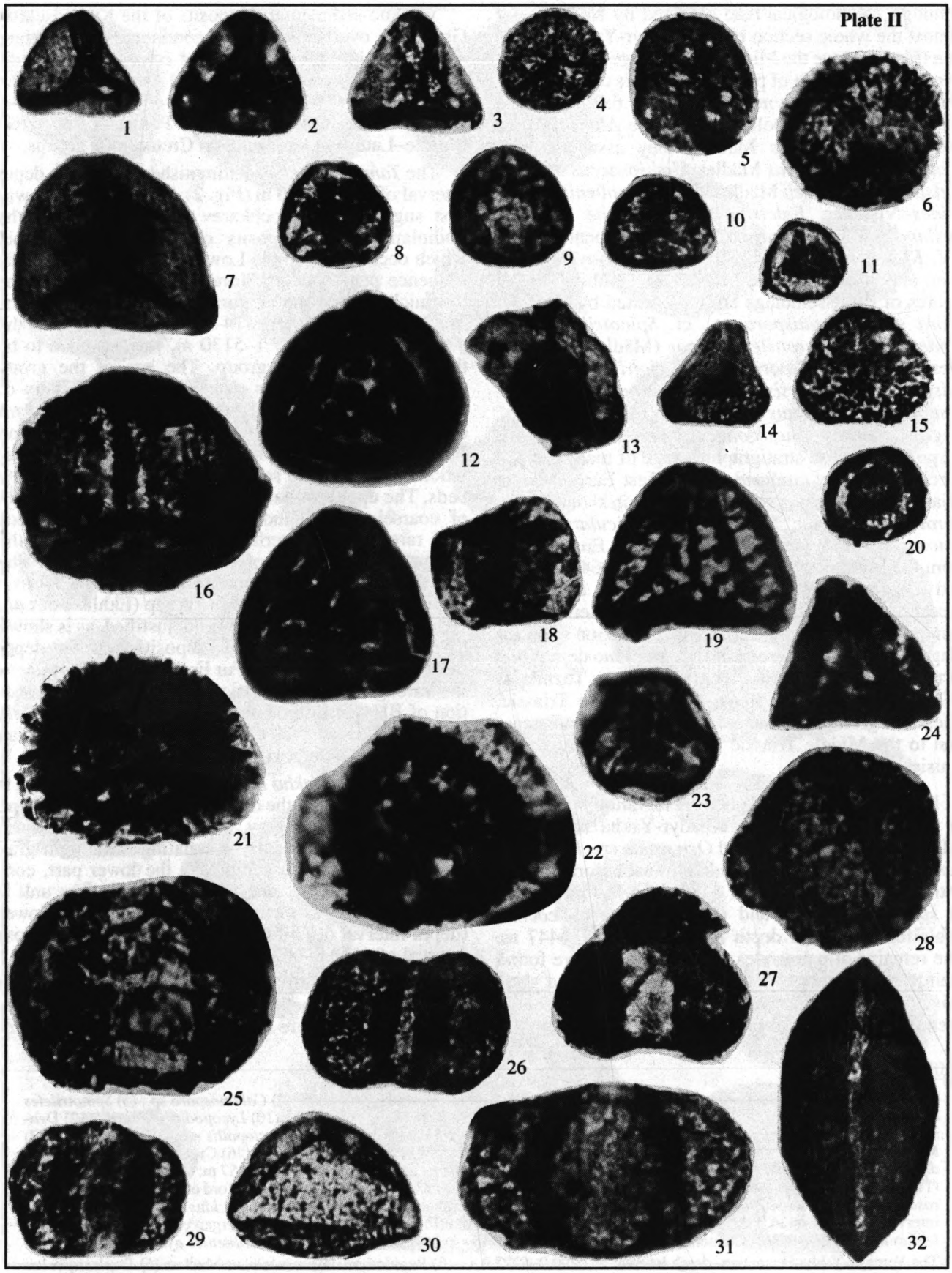
The *Tampei Group* is distinguished within the depth interval of 6421–5660 m (Fig. 2). The group name was first suggested by Bochkarev *et al.*, (1989) for the Ladinian–Carnian deposits of the Urengoi region, which occur between the Lower Triassic volcanogenic sequence penetrated by Borehole (BH) 46 within the Chernichnyi field and the Jurassic terrigenous deposits. The section of Borehole 414-P in the Urengoi area, the depth interval of 5287.1–5130 m, was assumed to be the stratotype of the group. The age of the group (Ladinian–Carnian) was determined on the basis of palynological data. We subdivided the group into two units, the lower Varenga-Yakha and upper Vityutino formations. The lower one is characterized by intercalation of fine-grained sediments and rare conglomerate beds. The upper formation is composed predominantly of coarse-grained sandstone and conglomerate beds with rare siltstone interlayers. In the TSB section, the Tampei Group is also of the two-unit structure. The idea that the third Pur Formation of the Middle Triassic age is present at the base of the group (Ekhlakov *et al.*, 1995; Nesterov *et al.*, 1995) is not justified, as is shown below. In terms of lithology, deposits from the depth interval of 6424–6011 m (Pur Formation) are similar to the Varenga-Yakha Formation from the stratotype section of BH 414 in the Urengoi area (Bochkarev *et al.*, 1989) and are characterized by the Ladinian–Carnian palynological assemblage.

The *Varenga-Yakha Formation* in the TSB section is distinguished within the depth interval of 6421–6050 m. It is represented by gray, dark-gray, and black mudstone and siltstone beds, alternating with light-gray sandstone and rare, especially in the lower part, conglomerate and coaly mudstone interlayers. The unit is characterized by spore–pollen assemblages in its lower (depth interval of 6409.6–6290.9 m) and upper (depth interval of 6120.8–6076 m) parts (Fig. 2, Plate II).

Pollen essentially dominates over spores in the assemblage from the lower part of the formation. Indicative pollen forms are bisaccate *Alisporites* and *Mono-*

Plate I. Khadyr-Yakha Formation, depth interval of 6490–6488 m: (1) *Punctatisporites* sp.; (2) *Calamospora* sp.; (5) *Spinotriletes echinoides* Mädler; (7) *Perotriletes minor* (Mädler) Clement-Westerhof; (8) *Spinotriletes* sp.; (10) *Lycopodiacidites* sp.; (12) *Densoisporites* sp.; (14) *Leptolepidites* sp.; (15) *Uvaesporites* sp.; (16) *Cyclotriletes* sp.; (21) *Colpectopollis ellipsoides* Visscher; (23) *Microchachryidites* sp.; (24) *Triadispora obscura* Scheuring; (25) *Caytonipollenites latus* Mädler; (26) *Cycadopites* sp.; (27) *Chordasporites* sp.; (29) *Falcisporites snophovae* Visscher; (30) *Alisporites* sp.; depth interval of 6447–6457 m: (4) *Apiculatisporites* sp.; (11) *Densoisporites* sp; (13) *Annulispora* sp.; (19) *Polycingulatisporites* cf. *densatus* (de Jersey) Playford et Dettmann; (20) *Umborosaccus* sp.; (22) *Alisporites* sp.; (33) *Colpectopollis ellipsoideus* Vischer; (34) *Taeniaesporites pellucidus* (Goubin) Balme; depth interval of 6412.9–6426.8 m: (17) *Nevesisporites limatulus* Playford; (18) *Aratrisporites saturni* (Thiergart) Mädler; (28) *Stellapollenites thiergartii* (Mädler) Clement-Westerhof; (31) *Sulcatisporites kraeuseli* Mädler; (32) *Striatoabietites aytugii* Visscher.

The Varenga-Yakha Formation. depth interval of 6290.9–6277.9 m: (3) *Rugulatisporites mesozoicus* Mädler; (6) *Duplexisporites* sp.; (9) *Cyclotriletes triassicus* Mädler.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sulcites associated with *Microchachryidites* sp., *Triadispora crassa* Visscher, *Heliosaccus dimorphus* Mädler, *Succinctisporites grandior* Leschik, *Striatoabietes aytugii* Visscher, and *Caytonipollenites*. Among spores of the assemblage, there are found diverse spiny forms *Osmundacidites* sp., *Apiculatisporites parvispinosus* (Lesch.) Schulz, and *Spinotriletes* sp. in association with *Calamospora* sp., *Punctatisporites* sp. (cf. *P. subcarpaticus* Pautsch), *Aratrisporites* sp., *A. scabratus* Klaus, *A. fischeri* Klaus, and *Leiotriletes* spp. Very rare are individual grains of *Duplexisporites* sp., *Kraeuselisporites* sp., *Cyclotriletes triassicus* Mädler, *Converrucosisporites* aff. *conferteornatus* Pautsch, and *Camarozonotriletes rudis* (Lesch.) Klaus. In the upper part of the formation, the abundance of *Leiotriletes* and spiny forms slightly increases, while that of *Calamospora* and *Punctatisporites* sharply decreases; *Anapiculatisporites telephorus* (Pautsch) Klaus, *Annulispora microanulata* de Jersey, and *Limbosporites lunndbladii* Nilsson are common taxa present here. In general terms of taxonomy, this spore–pollen assemblage is similar to the Varenga-Yakha assemblage from the stratotype section of BH 414 in the Urengoi area (Bochkarev *et al.*, 1989), where it is dated back to the Ladinian–Carnian stages, and to assemblage from the Chernyshovskii Formation of the Transurals (Tuzhikova, 1973). The fact that the Varenga-Yakha assemblage includes *Heliosaccus dimorphus*, the marker of the Ladinian Stage in western Europe (Mädler, 1964; Visscher and Brugman, 1981; Orlowska-Zwolinska, 1983), as well as *Anapiculatisporites telephorus* and *Succinctisporites grandior* abundant in the Ladinian–Carnian deposits of many regions, allows us to surely assign it to the Ladinian–Carnian time.

The fossil plant assemblage of the Varenga-Yakha Formation from the depth interval of 6060–6477 m is comprised of *Paracalamites* sp., *Neocalamites* sf. *carerrei* (Zeil.) Halle, *Neocalamites* sp., *Equisetites* sp., *Cladophlebis* sp., *Sphenopteris* sp., *Desmiophyllum* sp., *Carpolithes* sp. and is typical of the Ladinian–Carnian. The prevalence of horsetail stems representing genera *Paracalamites* and *Neocalamites* among plant fossils makes this assemblage similar to those from the stratotype (BH 414) and reference (BH 673) sections of the Urengoi area (Kirichikova and Travina, 1995).

The *Vityutino Formation* was recognized within the depth interval of 6050–5660 m. It is represented by alternating coarse-clastic rocks, mainly by medium- and coarse-grained sandstones (Fig. 2) intercalated with scarce conglomerate and gritstone interlayers as thick as 4 m at the base of the formation.

The spore–pollen assemblage was macerated from a siltstone interlayer (the depth interval of 6028–6010.9 m) at the base of the formation and studied by Kulikova. Among dominant *Leiotriletes* forms (*Concavisporites* and *Dictyophylides*; Plate II), there are *Duplexisporites gyratus* Playt. et Dett., *Anapiculatisporites spiniger* (Lesch.) Reich., *A. telephonus* (Pautsch) Klaus, *Annulispora microanulata* de Jersey, *Cingulizonates* sp., *Camarozonotriletes rudis* (Lesch.) Klaus, *Kyrtomisporites speciosus* Mädler, *K. laevigatus* Mädler, *Zebrasporites laevigatus* Schulz, and *Tigrisporites microrugulatus* Schulz. Common among the pollen are grains of *Alisporites australis* de Jersey, *Cycadopitys*, and *Chasmatosporites* (*C. apertus* Nilss., *C. hians* Nills.), accompanied by *Ovalipollis* sp., *Minutosaccus* sp., *Duplicisporites* cf. *gramlatus* Scher., and *Taeniaesporites rhaeticus* Schultz. Almost all species of the Vitytino assemblage are widespread and mark clear stratigraphic levels in many Triasic sections of Europe. For instance, species *Camarozonotriletes rudis, Anapiculatisporites spiniger*, and *Annulispora microanulata* are known from the Carnian–Rhaetian deposits in Germany and Poland (Schultz, 1967; Orlowska-Zwolinska, 1983); *Kyrtomisporites speciosus, Kyrtomisporites laevigatus, Zebrasporites laevigatus, Chasmatosporites apertus*, and *Taeniaesporites rhaeticus* occur in the uppermost Upper Triassic deposits (Rhaetian) of Europe. In the Tsvetkov Cape section of the Taimyr Peninsula, the first occurrence level of listed taxa is in the Carnian (Romanovskaya. 1989). This allows us to assign the Vityutino Formation to the Late Triassic (the Carnian–Norian).

Horsetail stems of genera *Paracalamites, Neocalamites* and *Equisetites* [*Paracalamites* sp., *N.* cf. *carerrei* (Zeil.) Halle, *Neocalamites* sp., *E.* cf. *turgaicus* (Vladim.) Kiritch., the depth interval of 5743–6050 m] and individual ferns mainly of the *Cladophlebis* genus are abundant among macroflora of the Vityutino Formation. This plant assemblage is basically similar to

Plate II. The Varenga-Yakha Formation, depth interval of 6488–6410.2 m: (1) *Dictyophyllidites mortonii* (de Jersey) Playford et Dettman. The same unit, depth interval of 6290.9–6277 m: (7) *Converrucosisporites* aff. *conferteornatus* Pautsch; (9) *Camarozonotriletes rudis* (Leschik) Klaus; (13) *Aratrisporites fischeri* (Klaus) Playford et Dettmann; (18) *Aratrisporites scabratus* Klaus; depth interval of 6277.9–6290.9 m: (2) *Concavisporites tumidus* Playford; (4) *Anapiculatisporites* sp., (11) *Stereisporites* sp.; (14) *Selagosporis mesozoicus* Schulz; (21) *Limbosporites lundbladii* Nilsson; (29) *Alisporites* sp.; (32) *Cycadopites* sp.; depth interval of 6080.8–6076.3 m: (5) *Apiculatisporites* sp.; (26) *Alisporites australis* de Jersey.

The Vityutino Formation, depth interval of 6038–6028.2 m: (3) *Dictyophyllidites mortoni* (de Jersey) Playford et Dettmann; (10) *Camarozonotriletes rudis* (Leschik) Klaus; (27) *Triadispora* sp.; (30) *Duplicisporites* sp.; depth interval of 6028.2–6019.4 m: (8) *Camarozonotriletes rudis* (Leschik) Klaus; (15) *Tigrisporites microrugulatus* Schulz; (16) *Kluperisporites baculatus* Schulz; (20) *Annulispora microannulata* de Jersey; depth interval of 6019.4–6010.9 m: (6) *Spinotriletes* sp.; (12) *Duplexisporites gyratus* Playford et Dettmann; (17) *Duplexisporites* sp.; (19) *Kyrtomisporis speciosus* Mädler; (22) *Cingulizonates* sp.; (23) *Zebrasporites laevigatus* (Schilz) Schulz; (24) *Kyrtomisporis laevigatus* Mädler; (25) *Taeniasporites rhaeticus* Schulz; (28) *Minutosaccus* sp.; the depth interval of 5769.5–5757.1 m, (31) *Alisporites* sp.

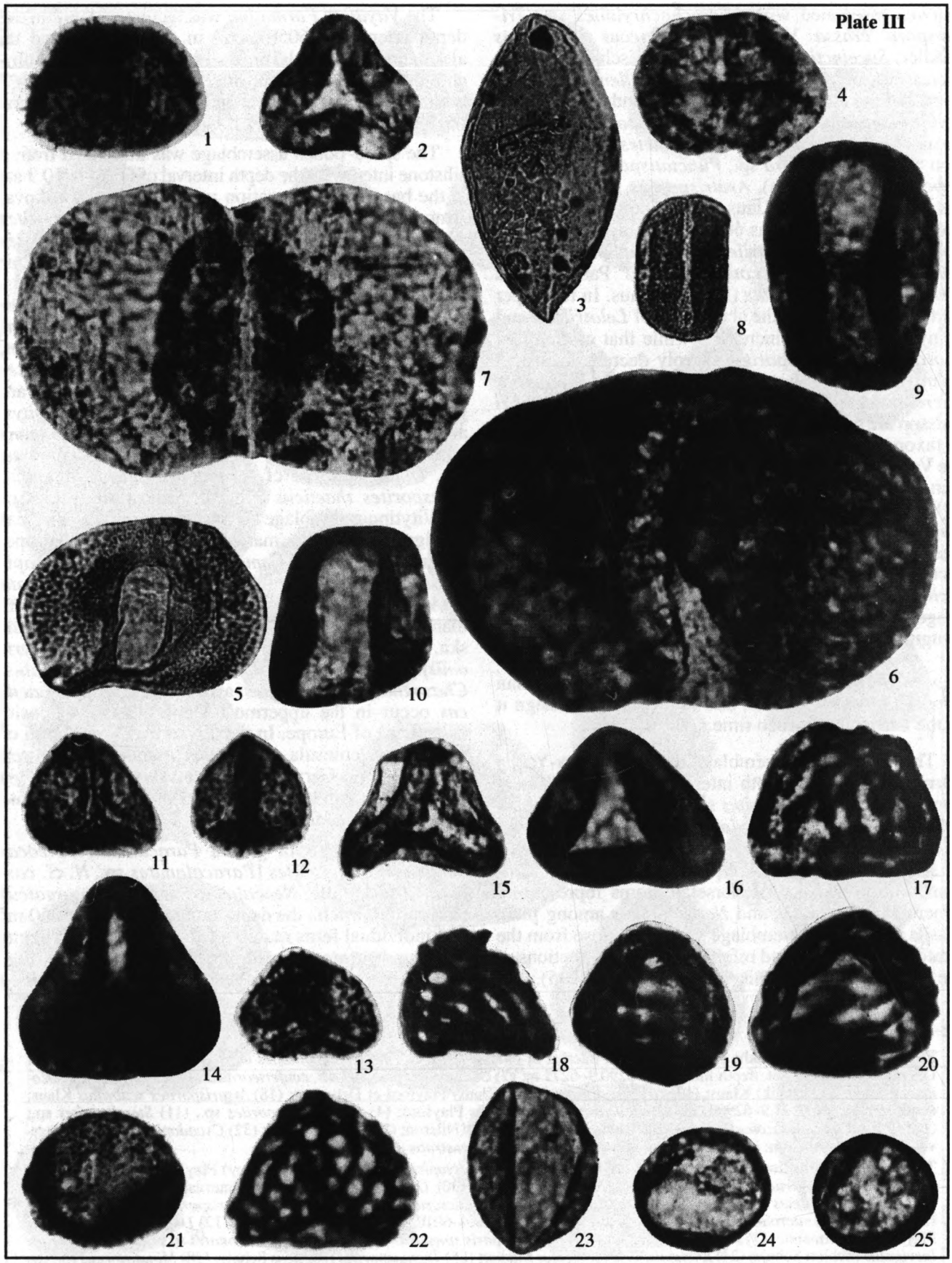
Plate III
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the Varenga-Yakha macroflora and has no difference from that of the Tampei Group penetrated by BH 414, 673, and 410 in the Urengoi area (Kirichkova and Travina, 1995). Moreover, the Tampei Group macroflora is identical in abundance of horsetail forms to the that of the Bulanash and Kalachevo formations of the eastern Urals (Kirichkova, 1990, 1993), and of the Nemtsov Formation of Tsvetkov Cape in Taimyr Peninsula (Mogucheva, 1982, 1984); all dated back to the Late Triassic (the Carnian–Norian).

JURASSIC

The *Zavodoukovskaya Group* comprises the Lower–Middle Jurassic deposits. According to Resolution of the Interdepartmental Stratigraphic Conference, it is subdivided (from the base upward) into the Beregovka, Yagel, Kotukhta and Tyumen formations.

The *Beregovka Formation* (the depth interval of 5660–5450 m), like in the hypostratotype section of BH 414 in the Urengoi area (Bochkarev *et al.*, 1989), is represented by coarse-grained sandstones with rare conglomerate and siltstone–shale interlayers (Fig. 2). The Early Jurassic palynological assemblage studied by Timoshina was recovered from siltstone interlayers of the lower part of the formation (the depth interval of 5660–5555 m). The assemblage (Plate III, images 1–10) includes abundant pollen of *Coniferales* and *Disaccites* associated with larger *Podocarpidites* grains *Quadraeculina* (*Q. limbata* Mal., rarely *Q. anellaeformis* Mal.), *Cycadopites, Ovalipollis, Chordasporites, and Platysaccus*. Identified among spores are species *Neoraistrickia* sp., *Duplexisporites* sp., *Leiotriletes* spp. (smaller forms), *Apiculatisporites, Osmundacidites* sp., and *Lycopodiumsporites* sp. Such a combination of taxons is characteristic of palynofloras from the lowermost Lower Jurassic of the Siberian paleofloral province, which are known from the Raspad Formation in Kuzbass, the lower Makarovo and Pereyaslovka subformations of the Kansk–Achinsk basin, and the lower beds of the Zimnee Formation of the Ust-Yenisei region; these palynofloras are dated back to the Hettangian–Sinemurian and, probably, early Pleinsbachian (Il'ina, 1985; Bochkarev *et al.*, 1989).

Among fossil plants, horsetail stems of the genus *Neocalamites* [*N.* cf. *carrerei* (Zeil.) Halle, *Neocalamites* sp.] and peculiar large-leaved Ginkgoales of the genus *Sphenobaiera* are encountered primarily in the Beregovka Formation. We have found similar divided leaves of *Sphenobaiera–Sphenobaiera magnifolia* Aksarin in a higher part of the section, namely, in the lower part of the Kotukhta Formation penetrated by BH 410 in the Urengoi area (Kirichkova and Travina, 1995). In general, the improverished plant assemblage from the Beregovka Formation is consistent in age with the dating based on palynological evidence. We recognized the boundary between the Triassic and Jurassic in the TSB section at a depth of 5660 m.

Poorly preserved microfauna, almost undeterminable even at the genus level, was found in the lower part of the Beregovka Formation. Species *Saccamina* sp. and *Ostracoda* sp. were encountered within the depth interval of 5592.9–5600 m, and only rare shell fragments were found within the depth interval of 5583.5–5591.5 m.

The *Yagel Formation* (the depth interval of 5450–5330 m) is represented in the TSB section by shale beds with siltstone and fine-grained sandstone interlayers. The formation yielded little or no fossil remains. Stem remains of horsetails *Neocalamites* sp. and *Equisetites* sp. (Fig. 2) are encountered in the middle part of the formation. Spores and pollen appeared to be of poor preservation, and only rare specimens of *Neoraistrickia* sp., *Cycadopites* sp., and *Disaccites* sp. were identified. The formation is assumed to be the Pleisbachian in age on the basis of its stratigraphic position between the Beregovka Formation (the lowermost Lower Jurassic) and the Kotukhta Formation, the lower part of which is assigned to the Toarcian.

The *Kotukhta Formation* (the depth interval of 5330–4610 m), which overlies the Yagel Formation, is represented in the lower part by irregularly alternating sandstone, siltstone, and mudstone beds (the depth interval of 5330–5120 m) intercalated with rare and thin (up to 10–14 cm) conglomerate and gritstone interlayers. A sequence of thin-bedded siltstone with subordinate siltstone and sandstone interlayers (probably the Togur unit) is distinguished within the depth interval of 5038–4980 m (Fig. 3). Upsection, polymictic small- to medium-grained thick sandstone beds alternate with black mudstone and siltstone interlayers bearing the irregularly distributed coaly material and plant remains (Ekhlakov *et al.*, 1991).

Plate III. The Beregovka Formation, depth interval of 5636–5623.5 m: (3) *Cycadopites medius* (Bolch.) Iljina; depth interval of 5600.1–5591.5 m: (1) *Spinotriletes* sp.; (4) *Chordasporites* sp.; (5) *Paleoconiferus rotundus* Odints.; (6) *Alisporites pergrandis* (Bolch.) Iljina; (8) *Quadraeculina anellaeformis* Mal.; (9, 10) *Quadraeculina limbata* Mal.; depth interval of 5591.5–5583.5 m: (2) *Leiotriletes* sp.; (7) *Podocarpidites* sp.

The Kotukhta Formation (the lower part), depth interval of 5046.8–5035.2 m: (11–13) *Dictyophyllum minutus* Mensh. et Timosh.; (16) *Leiotriletes* sp.; (21) *Osmundacidites* sp.; (22) *Klukisporites variegatus* Couper; (25) *Classopollis* sp.; depth interval of 5035.2–5021.9 m: (14) *Matonisporites* sp.; (18) *Duplexisporites* sp.; (24) *Classopollis* sp.; depth interval of 5021.9–5015.2 m: (19, 20) *Contignisporites problematicus* (Couper) Dor.; 23. *Chasmatosporites elegans* Nilsson; depth interval of 5015.2–5008.7 m: (15) Dipteridaceae gen. sp.; (17) *Duplexisporites* sp.

Magnification for all images is ×600.

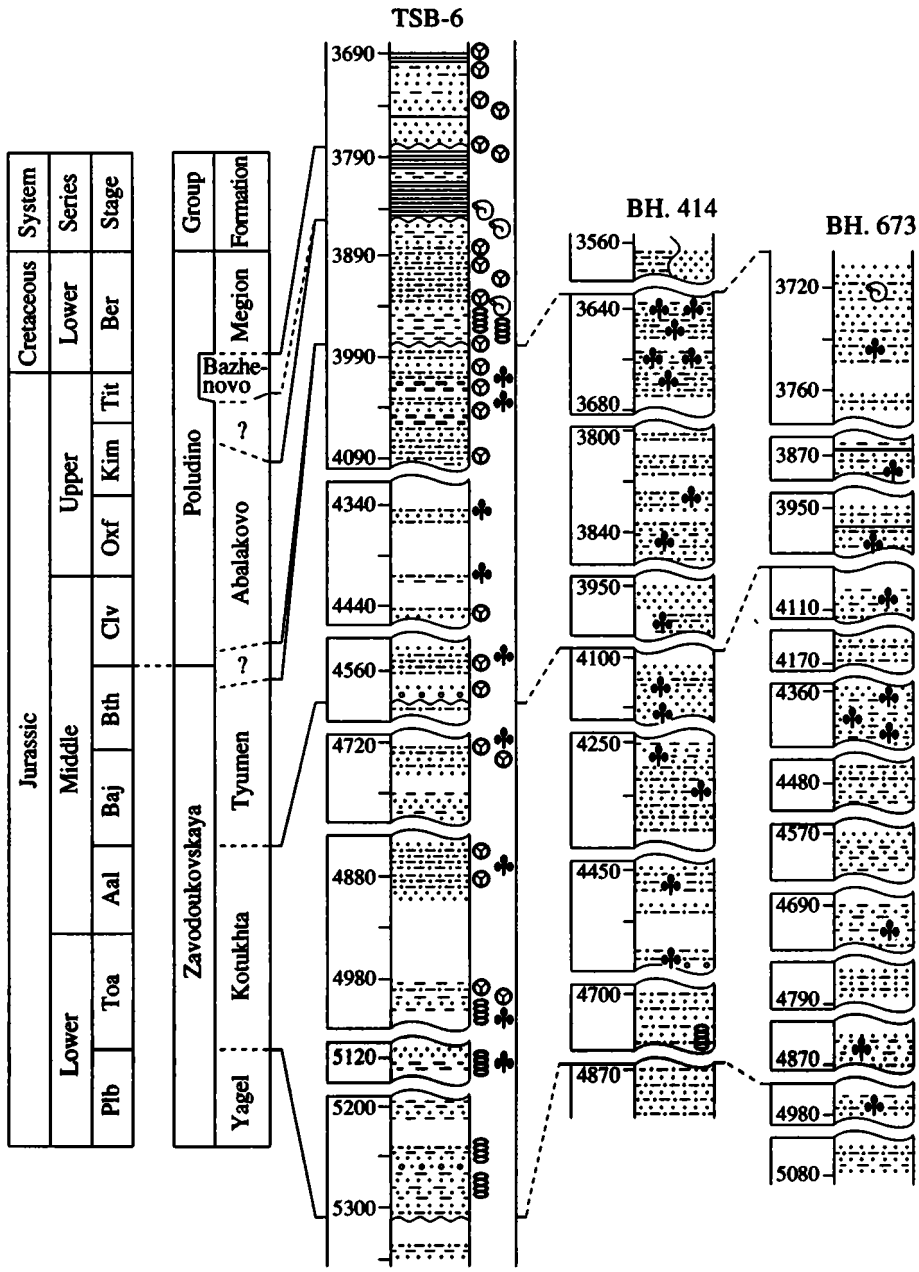

Fig. 3. Subdivision of the Jurassic and Lower Cretaceous deposits penetrated by TSB-6 and correlation with sections from reference boreholes of the Urengoi area (symbols as in Fig. 1).

Palynospectra are studied at different stratigraphic levels of the formation. In Timishina's opinion, the most representative is the palynological assemblage from the depth interval of 5046.8–4997 m. It includes low-abundant *Disaccites*, *Podocarpites*, larger *Cycadopites* grains, and individual specimens of *Quadraculina limbata* Mal. *Classopollis* pollen is abundant. Spores are diverse and abundant. Large- and medium-sized *Leiotriletes* forms dominate; *Contignisporites* sp., *Dictyophyllum* (spiny), and *Klukisporites variegatus* Couper are common; rare specimens represent Dipteridaceae gen. sp., *Tripatina variabilis* Mal., *Stereisporites* sp., *Cyathidites minor* Couper, *Microlepidites crassirimosus* Timosh., and *Osmundacidites* spp. (abundant in some sprectra). *Chasmatosporites* pollen was also encountered in places (Plate III, images 11–25). Timoshina identified a palynocomplex similar in composition, but more diversed in terms of taxonomy, within the lower half (the Togur Member) of the Sherkalinskoe Formation of the Krasnoleninskii arch in western Siberia and in the Ilanskii Formation of the Kansk basin; following Il'ina, she dated it back to the latest Early Jurassic (the Toarcian).

Although palynospectra from the depth intervals of 4873.0–4863.7 and 4734.5–4744.9 m are similar in composition to that described above, they lack *Classopollis* pollen and do not allow a confident conclusion on their Early Jurassic age. The palynospectrum from the depth interval of 4722.2–4734.5 m is rich in miospores. The spectrum includes larger *Podocarpidites* and *Quadraecula anellaeformis* Mal. characteristic of the Early Jurassic; encountered among abundant Coniferales is *Dipteralla oblatinoides* Mal. associated with spores of diverse *Leiotriletes* (including *L. karatauensis* Timosh.), *Osmundacidites* cf. *O. jurassicus* (K.-M. Kuz.), and *Lycopodiumsporites* (very abundant) of the Middle Jurassic type; also present are *Tripartina variabilis* Mal., *Stereisporites incertus* (Bolch.) Sem., *S. bujargiensis* (Bolch.) Schulz, *Hymenozonotriletes bicycla* (Mal.) Sach. et Fradk., *Cyathidites minor* Couper, Dipteridaceae, and a considerable amount of *Cycadopites* specimens characteristic of the Middle Jurassic. In its taxonomic composition, this assemblage correlates with palynospectra peculiar of the Kamala Formation of the Kansk basin and upper beds of the Sherkalinskoe Formation of the Krasnoleninsk arch, both identified by Timoshina as the Aalenian in age.

Scarce macroflora from the Kotukhta Formation is of a low diversity and concentrated in the Togur Member (the depth interval of 5046–5008 m, Fig. 3). *Neocalamites* sp., *Equisetites* sp., *Coniopteris murrayana* (Brongn.) Brongn., and *Cladophlebis* sp. represent this macroflora comparable with the fossil plant assemblage from lower beds of the Kotukhta Formation penetrated by BH 673 and 414 in the Urengoi area (Kirichkova and Travina, 1995). We consider the recovered flora as attributable to the latest Early Jurassic (presumably the Toarcian).

The upper part of the Kotukhta Formation in the TSB section (the depth interval of 4873–4722 m) yielded *Neocalamites* sp., *Equisetites lateralis* (Phill.) Phill., *E. beanii* (Bunb.) Sew., *Coniopteris* cf. *simplex* (L. et H.) Harris, *Cladophlebis nebbensis* (Brongn.) Nath., *Cladophlebis* sp., *Heilungia* ? sp., *Nilssonia* sp., *Carpolithes* sp., and *Pityophyllum* sp. The upper Kotukhta assemblage of the TSB section corresponds to palynospecra detected in the upper Kotukhta Subformation of the Urengoi area and in the upper Khudoseya Subformation of the Ob River and Tomsk regions, where they are attributed to the initial Middle Jurassic, arbitrarily to the Aalenian (Kirichkova and Travina, 1995). The boundary between the Early and Middle Jurassic in the TSB section is placed inside the Kotukhta Formation at the level of 4890 m.

Microfauna of the Kotukhta Formation is concentrated in its lower part within the depth interval of 4927–5297.9 m, where we detected such forms as *Saccamina* cf. *ampullacea* Schleifer, *Saccamina* sp., *Hyperammina* sp., *Ammodiscus* cf. *glumaceus* Gerke et Soss., *Ammodiscus* sp., *Glomospira* cf. *gordialis* (Parker et Jones), *Haplophragmoides* sp., *Ammobaculites* ex gr. *lobus* Gerke et Soss., *Ammobaculites* sp., *Trochammina* ? sp., *Lenticulina* (*Astacolus*) ex gr. *pallida* Schleifer var. *pallida* Schleifer, *Lenticulina* sp., *Ostracoda* ? sp., and also abundant but poorly preserved agglutinated foraminifers. According to the conclusion by Ovchinnikova, these microfossils are useless for stratigraphy. Their presence indicates only peculiar sedimentation environments during the Kotukhta time.

The *Tyumen Formation* overlaps with a minor unconformity the Kotukhta Formation. It is distinguished within the depth interval of 4610–3980 m (Fig. 3). In the lower part, the formation is composed of thick fine- to medium-grained sandstone beds alternating with thin interlayers of dark-gray, locally silty mudstone. The upper part of the formation shows the more regular alternation of sandstone, siltstone, mudstone, coaly siltstone, and thin coal beds.

Timoshina believes the age of palynospectrum characterizing the lower part of the Tumen Formation (the depth interval of 4564.2–4415.5 m) to be most likely the Bajocian. Abundant, poorly preserved *Disaccites* pollen and diverse spores are typical of the spectrum. Spores are represented by abundant and diverse *Leiotriletes*, *Osmundacidites* cf. *O. jurassicus* (K.-M.) Kuz., and *Lycopodiumsporites* spp. associated with less abundant *Hymenozonotriletes bicycla* (Mal.) Sach. et Fradk., *Tripartina variabilis* Mal., *Stereisporites* sp., *Cyathidites minor* Couper, *Microlepidites crassirimosus* Timosh., Dipteridaceae, and Middle Jurassic *Neoraistrickia rotundiformis* (K.-M.) Taras., *Dicksonia densa* Bolch., *Acanthotriletes* sp. (larger), and *Zonalopollenites dampieri* Balme; rare specimens of phytoplankton are also encountered. The assemblage is similar to those distinguished by Timoshina in deposits of the Borodin Formation in the Kansk basin and of the Tyumen Formation (its lower and middle parts) in the Tomsk, Ob River, and Krasnoleninsk arch regions. All these assemblages are attributed to the Middle Jurassic; arbitrarily, to the Bajocian.

The upper part of the Tyumen Formation (the depth interval of 4087–4002 m) is characterized by the palynospectrum dated by Timoshina as the Bathonian in age. The spectrum includes the following taxa: abundant but poorly preserved Coniferales and *Disaccites*; diverse and numerous *Leiotriletes*, *Osmundacidites*, and *Lycopodiumsporites*; less abundant *Neoraistrickia rotundiformis* (K.-M.) Taras., *Stereisporites incertus* (Bolch.) Sem., and *Tripartina variabilis* Mal.; characteristic *Klukisporites variegatus* Couper, *Matonisporites* sp., *Quadraeculina limbata* Mal., *Seiadopityspollenites* sp., and rare *Classopollis*; individual specimens of *Eboracia*. Scarce microphytoplankton specimens are persistent components. A similar assemblage was distinguished by Timoshina in the upper part of the Tyumen Formation in the Tomsk, Ob River, and Krasnoleninsk arch regions as characterizing the Bajocian microflora because of its similarity to that from the

lower part of the Malyshevka Formation (the Ust-Yenisei region), where it coexists with foraminiferal fauna (Rovnina, 1972; Bulynnikova and Yasovich, 1972).

Macroflora from the Tyumen Formation is scarce and confined mainly to its upper part. Its typical components are *Neocalamites* sp., *Equisetites lateralis* (Phill.) Phill., *Equisetites* sp., *Coniopteris* cf. *burejensis* (Zall.) Sew., *C.* cf. *simplex* (L. et H.) Harris, *C. depensis* E. Lebed., *Cladophlebis* cf. *williamsonii* (Brongn.) Brongn., *Raphaelia stricta* Vachr., *Sphenobaiera* sp., *Pityophyllum* sp., and *Desmiophyllum* sp. The combination of such forms as *Equisetites lateralis* (Phill.) Phill., *Coniopteris depensis* E.Lebed., *C.* cf. *burejensis* (Zall.) Sew., and C. cf. *simplex* (L. et H.) Harris allows a confident correlation of the upper Tumen macroflora of the TSB section with the Bajocian flora of the Urengoi, Ob River, and Tomsk regions (Kirichikova and Travina, 1995).

The *Poludino Group* is subdivided into the Abalakovo, Bazhenovo, and Megion formations.

The *Abalakovo Formation* (the depth interval of 3980–3840 m) of the marine genesis unconformably overlies the Tyumen Formation. Its lower part is composed of alternating dark-gray siltstone and mudstone beds with siderite lenses and pyrite inclusions. The upper part consists of gray polymictic sandstone overlapped by a thin (about 10 m) member of mudstone, siltstone, and sandstone beds with interlayers of greenish–gray to green glauconite siltstone (Ekhlakov *et al.*, 1991). The formation yielded macro- and microfauna and associated microphytofossils. Macrofauna was identified by Vyachkileva, Lebedev, and Turbina (ZapSibNIGNI). Ammonites *Quenstedtoceras (Soaniceras)* sp. indet, *Arctica* sp. (aff. *A. humiliculmina* Schur.), and *Bivalvia* gen. ind. (the late Callovian) were found within the depth interval of 3967–3928 m. *Dorsoplanites* sp. (the middle Volgian), along with bivalves *Buchia* cf. *mosquensis* (Bach) and *Inoceramus* ex gr. *vereschagini* Poch. (the middle Volgian), was encountered within the interval of 3825–3831 m.

The Callovian foraminiferal assemblage was studied by Ovchinnikova. Foraminifers are recovered from two intervals. The depth interval of 3976.6–3989.7 m yielded *Ammodiscus* sp., *Harlophragmoides* sp., *Recurvoides* cf. *scherkalyensis* Levina, *R.* cf. *tagaensis* Levina, and *Trochammina* sp. ind. Taxa from the interval of 3967–3976 m are *Saccamina* sp., *Ammodiscus uglicus chremcevae* Dain, *Haplophragmoides* sp., *Recurvoides scherkalyensis* Levina, *Ammobaculites* sp., *Dorothia* ? sp., *Lenticulina* sp. ind., and *Astacolus* ? sp.

The Callovian palynomorphs identified by Timoshina within the depth interval of 4000–3888.8 m include abundant pollen of Coniferales and *Disaccites*, along with *Sciadopityspollenites*, *Quadraeculina limbata* Mal., *Alisporites bisaccus* Rouse, and common *Classopollis* is (abundant in some spectra). Spores are more diverse and represented by *Leiotriletes* spp., *Osmundacidites jurassicus* (K.-M.) Kuz., *O.* spp., *Cyathidites minor* Couper, *Neoraistrickia rotundiformis* (K.-M.) Taras., *Lycopodiumsporites* sp., *Duplexisporites* sp., *Eboracia* sp., and *Klukisporites variegatus* Couper. Distinctive forms of the assemblage are large spores *Cyathidites australis* Couper, *Uvaesporites argentaeformis* (Bolch.) Schulz, and associate Selaginellacea and smaller *Acanthotriletes* spores. Microphytoplankton is persistently present, scarce in places, but often abundant and diverse, though poorly preserved. When compared, the Abalakovo assemblage from the superdeep borehole and those from the Vasyugan and Naunak formations of the Tomsk and Ob River regions display obvious similarity, although microphytoplankton occurs in the Vasyugan Formation only sporadically as single specimens.

The *Bazhenovo Formation* (the depth interval of 3840–3780 m) overlies the Abalakovo Formation. It is represented by gray to dark-gray mudstone and bituminiferous, pyritized shale. The Volgian (the lowermost part of the stage excluded) to early Berriasian age of the Bazhenovo Formation in the Urengoi area is substantiated in detail by ammonite, *Buchia*, and foraminiferal fauna (Braduchan *et al.*, 1986).

Microphytofossils studied by Fedorova were sampled from the upper part of the Bazhenovo Formation within the depth interval of 3778.3–3798.6 m (Fig. 3). Abundant specimens from this interval are represented by large cigar-shaped and globular particles probably of the algal origin. Individual prasinophytes (*Pterospermella* aff. ? *marginulata* Theod., *P.* aff. ? *griazevae* Theod.) and rare bisaccate pollen Coniferae gen. sp. indet. (including probably *Piceapollenites* sp. and problematic *Podocarpidites* sp.) were also identified among other specimens. The presence of algal remains of unknown taxonomy and rare transitory forms of Coniferae do not elucidate the age of enclosing rocks. It should be pointed out, however, that some prasinophyte species similar to those mentioned above were described by Fedorova (1990) from the Jurassic–Cretaceous boundary beds of the Russian platform in the Oka River basin (the central Volga region), where their stratigraphic range was identified as the middle–upper Volgian Substage of the Berriasian.

The *Megion Formation, Achimovka Member* (the depth interval of 3780–3690 m) is represented by light-gray to gray, polymictic sandstone with rare siltstone and mudstone interlayers. Fedorova identified microphytofossils from deposits in the depth interval of 3765.6–3698.2 m. The medium-sized and smaller pollen of *Pinuspollenites* spp., *Piceapollenites* spp., and *Podocarpidites* sp. dominate among Coniferae at this level. Most frequent spore taxa are *Leiotriletes*, *Cicatricosisporites minutaestriatus* (Bolch.) Pocock., problematic *C. australiensis* Pot. et Gell., and *C.* aff. *minor* (Bolch.) Pocock. Individual grains of *Sphagnumsporites* sp., *Densiosporites velatus* Weyl. et

Kireg., *Lycopodiumsporites* sp., *Osmundacidites* sp., problematic *Ishyosporites–Klukisporites* spp., ?*Lygodium* spp., and *Lophotriletes* sp. are almost persistently observed. The depth interval of 3765.8–3778.3 m yielded problematic *Aequitriradites* aff. *spinulosus* Delc. et Sprum. and *Foraminisporis* sp. in association with aff. *Tripartina* sp. and the more rare *Cupressacites* sp., *Caythonipollenites* sp., *Sciadopityspollenites* sp., and *Leptolepidites ? verrucatus* Coup. New-comers, such as problematic *Gnetaceaepollenites* sp., *Zonalapollenites* sp., *Leiotriletes* ? aff. *incertus* Bolch., *Selaginella* ? aff. *tenuispinulosa* Krasn., *Cicatricosisporites* ? *ludbrooki* Dett., *Lygodium* aff. *valanjinense* K.-M., *Lygodium* (*Impardecispora* ?) ? *triangulatum* E.Iv., problematic *Classopollis* sp., and *Stachycarpus* sp., appear in the depth interval of 3719.0–3731.9 m. Microphytoplankton from the lower part of the member is represented by scarce dinoflagellates: (?) *Muderongia* aff. *simplex* Alb., (??) *Gonyaulacysta* aff. *jurassica* (Defl.) Novik. et Sarj, (??) *Cribroperidinium* aff. *globatum* (Gitmez et Sarj) Helenes, *Tubotuberella* spp., *Gochteodinia* ? ex gr. *villosa* Norris, (?) *Dingodinium* sp., and (??) *Imbatodinium* aff. *kondratjevi* Vozzhen., as well as by individual prasinophyte species aff. (?) *Lancettopsis* sp., aff. (?) *Eupoikilofusa* sp., and *Pterospermella* ? aff. *parva* Theod.

Upsection, microphytoplankton appears to be more diverse in the composition of prasinophytes and dinocysts. Identified among them are *Tasmanites*, *Pterospermella* (including *P.* ? *helios* Sarj, *P.* ? aff. *magnicorpulenta* Theod., *P.* ? aff. *marginulata* Theod.) *Pleurozonaria* sp., *Pterosphaeridia* (forms with a large reticulum), *Eupoikilofusa* spp., *Paragonyalacysta* ? aff. *borealis* Brideauz, *P.* ?aff. *capillosa* Brideaux, *Cribroperidinium* sp., *Wallodinium krutzchi* Alb., Sentusidinium sp., *Hystrichodinium* aff. *voigthi* Alb., *Microdinium* ? *opacum* Brideaux, *Gochteodinia* ex gr. *villosa* Norris., and aff. (?) *Tubotuberella* sp. Microphytofossils from the Achimovka Member are similar in composition to their assemblages from the Berriasian deposits of central Siberia (Urdyuk-Khaya Cape, Paksa Peninsula; Il'ina, 1985) and arctic Canada (Brideaux and Fisher, 1976).

SUMMARY

The comprehensive study and analysis of biostratigraphy of Mesozoic deposits penetrated by TSB-6 resulted in the following:

(1) The complete Mesozoic section was first penetrated by a single borehole and revealed a succession of its lithological and biostratigraphic units.

(2) Biostratigraphy of the lower volcano-sedimentary part of the section was refined, and, following Kazanskii and co-authors (*Rezul'taty...*, 1995; Kazanskii *et al.*, 1995), the Krasnoselkup Group was diveded into three units: the Evo-Yakha (tentatively Permian), Korotchaevo (Induan–Olenikian of the Lower Triassic ?), and Khadyr-Yakha (Anisian of the Middle Triassic) formations. The boundary between the Permian and Triassic is arbitrarily placed at the level of 7250 m.

(3) The subdivision into formations and age of the sedimentary Tampei Group are refined. The two-unit structure of the group and its Ladinian–Carnian age in the TSB section are substantiated using correlation with the type and reference sections of the group in the Urengoi area boreholes (Bochkarev *et al.*, 1989; Kirichkova and Travina, 1995).

(4) Nomenclature of formations from the Lower Jurassic part of the TSB section is verified in accordance with the Resolutions of the Interdepartmental Stratigraphic Conference on West Siberia, Tyumen, 1990. A break in sedimentation at the end of the Late Triassic between the Vityutino and Beregovka formations is revealed.

(5) The Middle and Upper Jurassic Abalakovo and Bazhenovo formations of marine sediments and the lowermost Lower Cretaceous deposits (the Achimovka Member of the Megion Formation) was first characterized in detail by microphytofossils.

(6) The time succession of palynological assemblages characteristic of continental deposits is distinguished for the first time in the integral Mesozoic section. The revealed assemblages were used as a basis for extensive chronostratigraphic correlations within and outside the Urengoi region. The boundary between the Anisian and Ladinian in the TSB section is placed at a depth of 6421 m, and that between the Triassic and Early Jurassic, at a depth of 4890 m.

(7) The comprehensive biostratigraphic study on Mesozoic deposits penetrated by TSB-6 showed that their section can be considered as the reference one in the northern part of western Siberia.

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Biogeography of Planktonic Foraminifers in the Paleocene and Eocene North Pacific

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Received April 22, 1997

Abstract—Changes in species diversity, taxonomy, and structure of late Paleocene–Eocene planktonic foraminiferal assemblages are discussed using the results of the analysis of their distribution in the North Pacific region along the profile from the Mid-Pacific Mountains in the south to northern Kamchatka in the north. It is shown that, during the late Paleocene and early Eocene, the North Pacific was populated by a foraminiferal assemblage dominated by representatives of the equatorial–tropical (*Morozovella*) and tropical–subtropical (*Acarinina*) genera. Their role slightly decreases only in northern Kamchatka. In the middle Eocene, planktonic foraminifers of the North Pacific showed a distinct latitudinal differentiation, and this situation was particularly well-expressed in the late Eocene. Similar tendency is also characteristic of the southern high-latitude zone of the Pacific. As a whole, changes in the taxonomic composition of planktonic foraminiferal assemblages, their structure, and distribution in the study region during the late Paleocene and Eocene are similar to those established previously in the Atlantic.

Key words: *late Paleocene, Eocene, planktonic foraminifers, biogeography, latitudinal differentiation, North Pacific.*

INTRODUCTION

The Paleogene period in the earth's history was a time of dramatic oceanographic reorganization, when the haline circulation characteristic of the Cretaceous ocean was replaced by that of the thermohaline type. The oceanographic reorganization occurred in parallel with the global cooling, which commenced in the middle Eocene and continued during the entire Cenozoic. There were periods of relative warming alternating with periods of sharp decrease in the temperature of surficial and bottom waters against the background of gradual cooling global trend. Oceanographic and climatic events were recorded in changes of the composition and diversity of biota, particularly of planktonic microorganisms, foraminifers included. These processes were most distinct in calcareous planktonic assemblages inhabiting high-latitude areas, where they responded to relative temperature fluctuations not only by changes in the assemblage structure, i.e., by changes in proportions of thermophilic and cold-resistant species, but also by changes in species diversity because of the disappearance and appearance of particular species. Therefore, the study of planktonic foraminiferal taxonomy and the dynamics of spatial and temporal changes in assemblages of these fossils is extremely important from the viewpoint of both biogeographic and climatic interpretations.

Until recently, Paleogene planktonic foraminiferal assemblages of the high-latitude areas of the Pacific were virtually unknown; therefore, their biogeographic affinity was poorly understood. The high-latitude sections of Paleogene deposits in the North Pacific were drilled only at the Schatsky and Hess rises, which are located nowadays near the northern boundary of the northern subtropical zone. As for the boreal zone of the Pacific, planktonic foraminifers were known there only from Paleogene deposits of the Meiji Guyot penetrated at Site 192 in Leg 19 of the D/V *Glomar Challenger* of the Deep Sea Drilling Project, where they are characterized by low abundance and poor preservation (Greager *et al.*, 1973). Paleogene sections composed mainly of marine terrigenous and volcanogenic–terrigenous deposits with the planktonic foraminiferal fauna are also known from the western continental margin of the Bering Sea (Krasheninnikov *et al.*, 1988; Ben'yamovskii *et al.*, 1992; *Opornyi razrez...*, 1994; Ben'yamovskii and Gladenkov, 1996).

Relatively complete sections of Paleogene deposits in the high-latitude area of the Pacific, with diverse assemblages of calcareous planktonic microorganisms, foraminifers included, were first drilled at the Obruchev Rise in Leg 145 of the D/V *JOIDES Resolution* of the Ocean Drilling Program (Rea *et al.*, 1995). Paleogene sediments are penetrated at Site 883 (51°11,908'N; 167°20,128'E; water depth 2385 m) on the summit of the rise and Site 884 (51°27,026'N; 168°20,228'E; water depth 3825 m) drilled at the foot of its northeastern slope.

Planktonic foraminifers are present in both sections, being more diverse, abundant, and better-preserved at

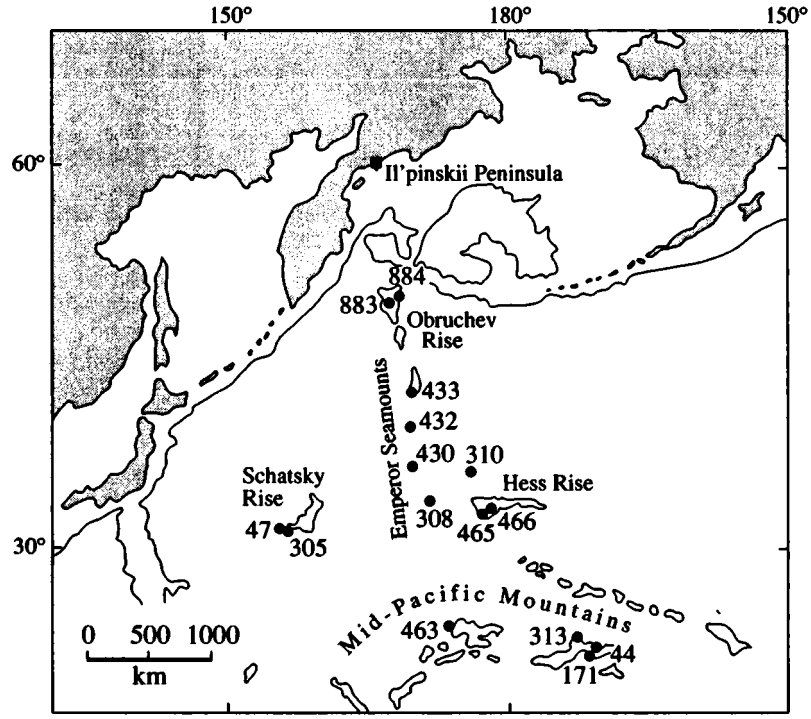

Fig. 1. Location deep-sea drilling sites in the North Pacific.

the relatively shallow-water Site 883, where they occur, though very unevenly, throughout the entire Paleogene section. At the deeper Site 884, the impoverished planktonic foraminiferal assemblage is present only in the middle Eocene interval. The diversity, abundance, and preservation degree of planktonic foraminifers from the Obruchev Rise were considered earlier (Basov, 1995, 1997). Based on planktonic foraminifers and nannofossils, the Paleogene section was subdivided into the upper Paleocene, lower Eocene, middle Eocene, and upper Eocene(?)–lower Oligocene stratigraphic units.

Previously, it was shown (Krasheninnikov *et al.*, 1988; Ben'yamovskii and Gladenkov, 1996) that periods of warming in the Paleogene were marked by migration of tropical–subtropical foraminiferal species into high latitudes of the Pacific region, which generally were characterized by relatively cold-resistant planktonic foraminiferal assemblage. The aim of the paper is to elucidate the structure of planktonic foraminiferal assemblages at different latitudes of the Pacific region and its changes throughout the Paleogene.

MATERIALS AND METHODS

Data on the composition and distribution of planktonic foraminifers in Paleogene deposits drilled at the Obruchev Rise (Sites 883 and 884) in Leg 145 of the D/V *JOIDES Resolution* of the Ocean Drilling Program served as a basis for the analysis of their biogeography in the high-latitude areas of the Pacific. Paleogene sediments from these sections were sampled for the foraminiferal analysis (69 and 145 samples, respectively). Samples were treated using the conventional procedure to obtain residues for the study under the microscope. Planktonic foraminifers were registered almost in all the samples from Hole 883 and only in 20 samples from Hole 884.

To compare planktonic foraminiferal assemblages from the boreal zone with their coeval associations from lower and southern high latitudes of the Pacific, the published data were used (Jenkins, 1971; Luterbacher, 1975; Butt, 1982; Krasheninnikov, 1982; Krasheninnikov and Basov, 1986; *Opornyi razrez...*, 1994; and others). Paleogene foraminifers from particular site areas were studied by different scientists. For instance, data on planktonic foraminifers from Paleogene sediments of the Schatsky Rise can be found in several works (Luterbacher, 1975; Krasheninnikov, 1982; Corfield and Shackleton, 1988). Taking into consideration that different researchers often treat some species in different ways, data on foraminiferal taxonomy were taken from the work with most complete description of their assemblages (in the above-considered case, from Krasheninnikov, 1982). Figure 1 demonstrates the position of deep-sea drilling holes and main on-land Paleogene sections used for the biogeographic analysis of foraminiferal assemblages.

For the analysis of the planktonic foraminifera composition, their taxa were broken down into groups of

relatively thermophilic (tropical and subtropical) and cold-resistant species using the classification proposed by Boersma and colleagues (Boersma and Premoli Silva, 1983, 1991; Boersma *et al.*, 1987; Premoli Silva and Boersma, 1989). Based on the analysis of planktonic foraminifera distribution in Paleogene sediments drilled in all climatic zones of the Atlantic, these researchers established that the late Paleocene–early Eocene foraminiferal assemblage was universally dominated by representatives of *Morozovella* and *Acarinina* genera. First of these genera is an equatorial–tropical taxon, whereas the other is a tropical–subtropical one. They were accompanied by temperate species of *Guembelitria, Chiloguembelina*, and *Planorotalites* genera, as well as by low-conical subbotinids and tenitellids characteristic of high-latitude areas. In the middle Eocene, the tropical areas were mainly populated, along with *Morozovella* and *Acarinina* species, by representatives of *Hantkenina, Globigerinatheka*, and *Turborotalia* (*T. cerroazulensis* group). Species of *Truncatulinoides* and *Orbulinoides* genera, which occur permanently in Paleogene sediments of the Pacific tropical zone (Krasheninnikov *et al.*, 1988), are evidently also indicators of warm-water conditions in the middle Eocene time. Representatives of these genera populated also the temperate zone, where they were dominant along with pseudohastigerinids and low-conical subbotinids.

Indicators of cold high-latitude waters were low-conical subbotinids, tenuitellids, and representatives of *Chiloguembelina, Globorotaloides*, and *Catpsydrax* genera. According to the above-mentioned authors, the late Eocene tropical ocean was characterized by a wide distribution of *Hantkenina, Globigerinatheka, Globoquadrina, Paragloborotalia, Dentoglobigerina, Cassigerinella*, and *Turborotalia* species. Many of them were widespread in moderate latitudes as well. Pseudohastigerinids, low-conical subbotinids, and large globigerinds were the dominant element of foraminiferal assemblages in these areas. High-latitude waters were mainly populated by species of *Chiloguembelina, Globorotaloides, Globigerina, Catapsydrax*, and *Tenuitella* genera (Boersma and Premoli Silva, 1991).

The most reliable interpretations of paleoenvironments are obtained by analyzing the quantitative characteristics of particular species and peculiarities in the assemblage structure. Unfortunately, virtually all the publications dealing with Paleogene planktonic foraminifers of the Pacific lack quantitative data. Usually, they contain data on the relative content of species expressed in categories "abundant," "common," "few," and others. Nonetheless, foraminiferal studies in different climatic zones of the Atlantic showed that biogeographic interpretations are virtually independent of what characteristics (quantitative or relative) are analyzed (Boersma and Premoli Silva, 1991). It should be noted that, under intense postdepositional dissolution, proportions of separate species in taphocoenoses can be strongly distorted as compared with those in biocoenoses, whereas the taxonomic composition could remain unchanged. Most of the Paleogene sections drilled in the North Pacific are fragmentary and their subdivision is different in detail, which often makes their comparison, using zonal units, impossible. Sometimes, because of the different completeness of sections, one should compare assemblages *a fortiori* different in their quantitative composition, which naturally can negatively affect interpretation based on the species diversity in certain time intervals or climatic zones. It is clear, however, that proportions of relatively thermophilic and cold-resistant elements in the assemblages should remain unchanged regardless of their preservation degree.

Data on the taxonomic composition of late Paleocene and Eocene planktonic foraminifers and their distribution in Paleogene deposits of different latitudinal belts of the northwestern Pacific are given in the table.

RESULTS AND DISCUSSION

In general, planktonic foraminiferal assemblages from Paleogene sediments of the Obruchev Rise show a mixed composition. They include both the typical thermophilic (tropical) forms and the species peculiar of temperate and higher latitudes. Proportions of these species vary with time in responce to global cooling in the Cenozoic and, probably, local peculiarities in the climatic and hydrological environments.

In order to trace these variations and define, following the above-mentioned classification (Boersma *et al.*, 1987; Boersma and Premoli Silva, 1991), the biogeographic affinity of Paleogene assemblages of planktonic foraminifers from the Obruchev Rise, diagrams demonstrating relations of thermophilic to cold-resistant species were compiled separately for the late Paleocene, early Eocene, middle Eocene, and late Eocene.

Figure 2 shows the dynamics of changes in the species diversity and structure of planktonic foraminiferal assemblages in the Obruchev Rise area during the late Paleocene–Eocene. The upper Paleocene deposits contain an extremely impoverished foraminiferal assemblage. Taking into consideration that the Obruchev Rise of the late Paleocene time was situated at least by 20° south of its modern position (Rea *et al.*, 1995), and the surface water temperature, according to isotopic studies (Pak and Miller, 1995), was high in the North Pacific, this impoverishment of the foraminiferal assemblage was most likely a result of unfavorable conditions for either burial or dwelling in the area at that time (Basov, 1997). It is also probable that it was caused by a combination of both these factors. Indeed, such a composition prevents a reliable paleoecological interpretation. Nonetheless, it is noteworthy that three out of four (or 75%) late Paleocene species discovered

Planktonic foraminifers from Paleogene deposits of the North Pacific region

Species	I	II	III	IV	V	VI
Acarinina						
acarinata	+	+	+			+
angulosa		+				
apressocamerata		+				
aquiensis			+			
aspensis	+	+	+			
berggreni		+				
broedermanni	+	+				
bullbrooki	+	+	+	+	+	
camerata	+	+				
carinata					+	
clara			+			
collactea		+				
cuneiacamerata		+				
decepta		+				
esnaensis		+	+			+
graveli		+				
intermedia						+
interposita	+	+				
matthewsae				+	+	
mckannai		+	+	+	+	+
multiloculata		+				
nitida		+			+	
pentacamerata	+	+	+		+	+
primitiva		+			+	
pseudotopilensis	+	+	+	+		+
rotundimarginata	+		+			
rugosoaculeata					+	
soldadoensis	+	+	+		+	+
strabocella		+				
subsphaerica					+	+
tadjikistanensis djanensis	+		+			
tribulosa		+				
triplex	+	+	+		+	
wilcoxensis		+			+	
Catapxydrax						
echinatus	+					
howei	+					
martini					+	
perus	+					
unicavus					+	
Chiloguembelina						
wilcoxensis			+	+		

Table. (Contd.)

Species	I	II	III	IV	V	VI
Cribrohantkenina						
inflata	+					
Dentoglobigerina						
galavisi	+				+	
Eoglobigerina						
trivialis		+	+			+
varianta		+	+			+
Globigerina						
angustiumbilicata					+	
increta cea	+					
praebulloides	+				+	
pseudovenezuelana	+					
tripartita	+					
Globigerinatheka						
barri	+	+				
index	+		+		+	+
kugleri	+		+			
semiinvoluta	+	+				
tropicalis	+	+				
Globorotalia						
imitata		+	+			+
reissi						+
Globorotaloides						
carcosellensis					+	
suteri	+				+	
taroubaensis		+				
testarugosa					+	
turgidus	+					
Hantkenina						
alabamensis	+					
lehneri	+					
longispina	+					
suprasuturalis	+					
Morozovella						
acuta		+	+			
aequa	+	+	+	+		+
angulata		+	+	+		
apanthesma		+				
aragonensis	+	+	+		+	
caucasica	+	+	+	+	+	
conicotruncata		+	+	+		
convexa		+	+			+
crater		+				
edgari		+				
elongata		+		+		+

Table. (Contd.)

Species	I	II	III	IV	V	VI
formosa formosa	+	+	+			
formosa gracilis	+	+		+	+	
hispicidaris		+	+			
kubanensis		+	+			
laevigata		+	+			+
lehneri	+	+				
lensiformis		+	+	+	+	
marginodentata	+	+		+		
marksi		+	+			
naussi		+				
nicoli						+
occlusa		+	+			
parva		+				
pasionensis		+	+			
perclara						+
quadrata		+	+			
quetra	+	+	+		+	
simulatilis				+		
spinulosa	+	+		+		
subbotinae	+	+		+	+	+
tortiva		+				
trichotrocha		+				
velascoensis		+	+	+	+	
wilcoxensis	+	+				+
Muricoglobigerina						
senni	+	+	+	+	+	
Orbulinoides						
beckmanni	+					
Planorotalites						
compressus		+		+		
ehrenbergi		+	+			
planoconicus		+				
pseudomenardii		+	+	+		
pseudoscitulus			+			
pusilla		+	+	+		
Pseudohastigerina						
micra	+	+			+	+
wilcoxensis	+	+		+		+
Subbotina						
angipora						+
angiporoides					+	
bacuana		+				+
chascanona		+				
compressaformis	+	+	+			+

Table. (Contd.)

Species	I	II	III	IV	V	VI
corpulenta	+	+				
eocaena		+			+	+
eocaenica					+	+
hagni					+	
hevensis						+
higginsi		+	+	+		+
inaequispira	+	+				+
linaperta					+	
lozanoi					+	
nana	+	+	+	+		
pileata		+				
prolata		+				
pseudobullodes		+		+		
pseudoecoaena compacta	+					
pseudoecoaena pseudoecoaena						+
pseudoecoaena tribulosa						+
posttriloculinoides	+					+
quadritrioculinoides		+				+
rotundaenana						+
triloculinoides	+	+		+		+
velascoensis		+	+		+	+
yeguaensis						+
zuensis						+
Tenuitella						
gemma					+	
Truncarotaloides						
rohri	+	+			+	
topilensis	+	+				
Turborotalia						
bolivariana	+					+
boweri		+	+		+	+
centralis	+					
cerroazulensis	+	+				
cocoaensis	+					
frontosa	+				+	+
griffinae					+	

Note: Roman numbers in columns: (I) Mid-Pacific Mountains; (II) Schatsky Rise; (III) Hess Rise; (IV) Emperor Seamounts; (V) Obruchev Rise; (VI) Koryak–Kamchatka region.

there belong to the thermophilic genera *Morozovella* and *Acarinina*.

The early Eocene planktonic foraminiferal assemblage of the area in question is significantly more diverse as compared with the late Paleocene one (21 species). The share of thermophilic species in this assemblage is higher than in the latter. Together, species of the genera *Morozovella* and *Acarinina* (7 and 11 species, respectively) make up about 85%. The remaining 15% are representatives of low-conical subbotinids characteristic of higher latitudes. Noteworthy is the insignificant, like in the late Paleocene assemblage, share of species, which could be indicators of climatic

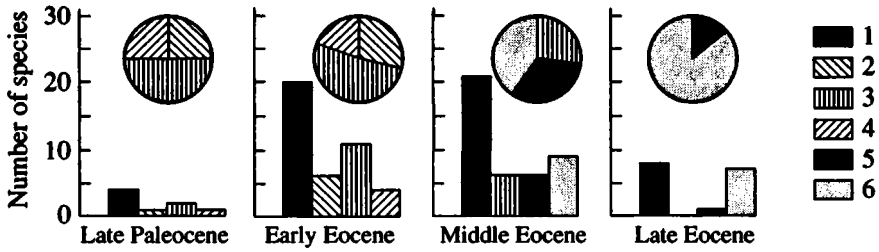

Fig. 2. Changes in the species diversity and structure of planktonic foraminiferal assemblages from Paleogene deposits of the Obruchev Rise (Sites 883, 884): (1) total species number; (2, 3) tropical *Morozovella* (2) and *Acarinina* (3) forms; (4, 5) temperate forms of (4) mainly *Subbotina* genus (late Paleocene–early Eocene), (5) *Globigerinatheka, Hantkenina, Pseudohastigerina, Truncorotaloides, Turborotalia, Dentoglobigerina* genera (middle–late Eocene); and (6) cold-resistant taxa (*Catapsydrax, Globorotaloides, Globigerina*).

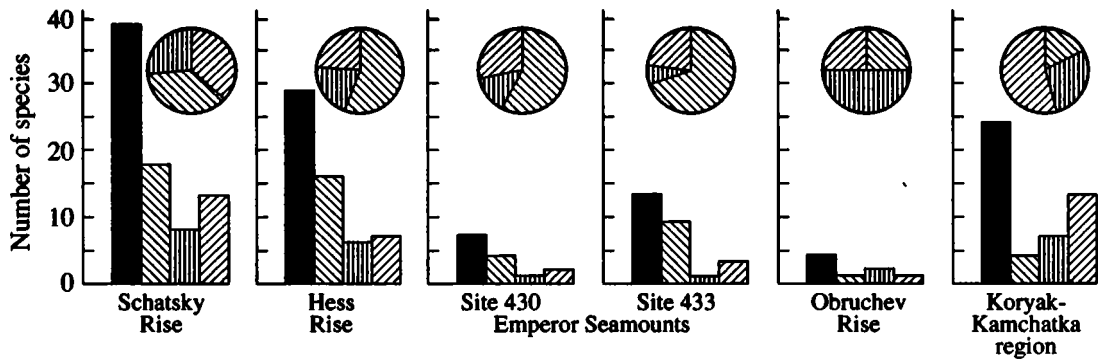

Fig. 3. Changes in the species diversity and composition of late Paleocene planktonic foraminifers in the North Pacific region (symbols as in Fig. 2).

conditions peculiar of modern temperate and higher latitudes.

The situation noticeably changed in the middle Eocene. Despite the fact that the total diversity of planktonic foraminifers remained virtually unchanged at that time (22 species), the proportion of thermophilic and cold-resistant species became different. The share of cold-resistant species (representatives of the genera *Subbotina*, *Catapsydrax*, and *Globorotaloides*) increases to about 40%, whereas that of species belonging to the typical thermophilic genera, such as *Acarinina* and *Globigerinatheka*, decreased to 28 %.

The middle Eocene was marked by the appearance of species, which, according to the accepted classification, were characteristic of temperate (subtropical) environments. These are *Turborotalia frontosa*, *T. boweri*, *T. griffinae*, and *Subbotina lozanoi* representing approximately one third of the total taxonomic composition of the foraminiferal assemblage.

Thus, thermophilic (tropical and subtropical) species in the middle Eocene assemblage of the Obruchev Rise made up slightly more than 50% in total. The remaining part of the assemblage was represented by cold-resistant (cosmopolitan) forms.

Even more drastic changes in planktonic foraminiferal assemblages of the region occurred in the late Eocene(?)–early Oligocene. This period was characterized not only by a sharp diversity decrease, but also by the most significant reorganization of the assemblage structure. All seven planktonic foraminiferal species recorded in sediments of this age (genera *Catapsydrax*, *Globorotaloides*, *Globigerina*, and *Tenuitella*) are characteristic of cold-water conditions of high latitudes (Boersma and Premoli Silva, 1991).

In order to understand the position of planktonic foraminiferal assemblages from Paleogene deposits of the Obruchev Rise in the succession of climatic zones of that time, changes in the proportions of thermophilic and cold-resistant species should be traced when moving from low to higher latitudes. For this purpose, Paleogene foraminiferal assemblages were analyzed along the south-north profile from the Mid-Pacific Mountains (Sites 44, 171, 313, 463 of the Deep Sea Drilling Project; modern position about 20°N) in the south via the Schatsky Rise (Sites 47, 305; 32°–33°N), Hess Rise (Sites 310, 465, 466; 35°–38°N), Emperor Seamounts (Site 308; 34°59'N; Site 430; 37°59'N; Site 432; 41°20'N; Site 433; 40°45'N), and Obruchev Rise (Sites 883, 884; 51°N) to northern Kamchatka (approximately 60°N) in the north. Thereafter, plots demonstrating proportions of thermophilic and cold-resistant species in these areas were compiled using the abovementioned classification (Boersma and Premoli Silva,

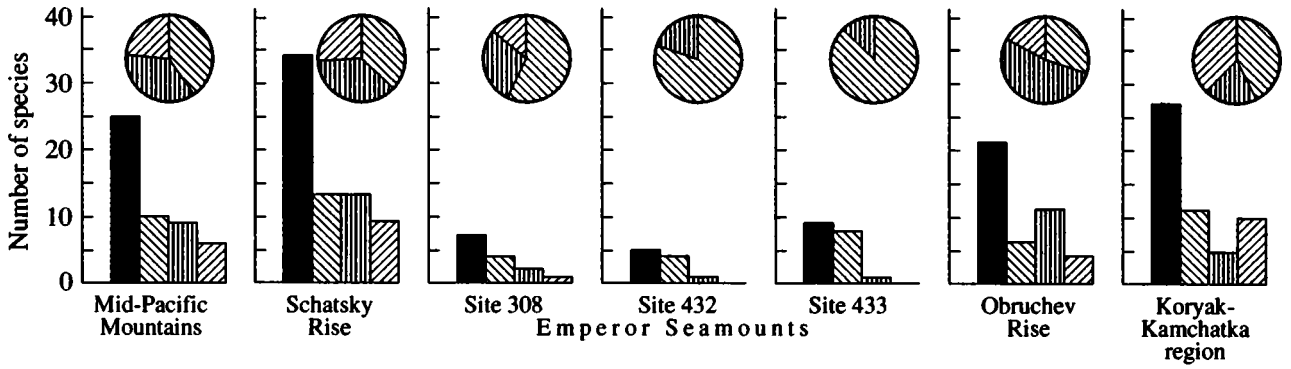

Fig. 4. Changes in the species diversity and composition of early Eocene planktonic foraminifers in the North Pacific region (symbols as in Fig. 2).

1991). Such plots were compiled separately for the late Paleocene, early, middle, and late Eocene time.

Figure 3 shows changes in the structure of the late Paleocene planktonic foraminiferal assemblage that are observed along the profile from the Schatsky Rise (in the Mid-Pacific Mountains area, Paleocene deposits are eroded) in the south to the Koryak–Kamchatka region in the north. Diagrams show that the total species diversity of foraminifers changes in this direction. In the Schatsky and Hess rises, 39 and 29 species, respectively, were registered (Krasheninnikov, 1982). In sites located along the Emperor Seamounts, where the Paleogene section is composed of shallow-water sediments (Hagn *et al.*, 1980), the foraminiferal assemblages are substantially impoverished, as compared with that from the Schatsky Rise, and include 7 species in the southernmost area of the ridge (Site 430) to 13 species in northern Site 433 (Butt, 1980; Krasheninnikov, 1982). The least diverse foraminiferal assemblage (4 species) is encountered at the Obruchev Rise (Site 883), where, as was mentioned, the local conditions were most likely unfavorable for dwelling or burial. A rather diverse assemblage of late Paleocene foraminifers is discovered from the well-studied section of the Il'pinskii Peninsula in the Koryak–Kamchatka region, where it consists of 24 species (Krasheninnikov *et al.*, 1988).

Despite such a difference in the diversity, substantially depending on both the state of knowledge and local dwelling or burial conditions, the structure of foraminiferal assemblages remains more or less stable in all the mentioned regions except the latter. In the Schatsky and Hess rises, Emperor Seamounts, and Obruchev Rise, the summary abundance of thermophilic species of the genera *Morozovella* and *Acarinina* makes up 67, 76, 71–73, and 75%, respectively. Only in the Koryak–Kamchatka region, the share of these genera decreases to 46%. The remaining part of the assemblage is represented by low-conical *Subbotina*, *Tenuitella*, and *Planorotalites* forms.

A similar, but even more expressive picture is characteristic of the early Eocene. The early Eocene assemblages of planktonic foraminifers are known in the North Pacific region from the Mid-Pacific Mountains, Schatsky Rise, southern Emperor Seamounts, Obruchev Rise, and from the Koryak–Kamchatka region. Similarly to the underlying deposits, the lower Eocene sediments show significant variations in the diversity of foraminiferal assemblages along the considered profile (Fig. 4).

In the Mid-Pacific Mountains and Schatsky Rise, where the early Eocene assemblages of planktonic foraminifers include 25 and 33 species, respectively, more than three quarters of the species are represented by *Acarinina* and *Morozovella* forms, whereas the relatively cold-resistant species make up only 24 and 33%, respectively. In the southern extremity of the Emperor Seamounts (Site 308), the share of these species decreases to 14% of the assemblage consisting of 7 species; the dominant role also belongs to morozovellids and acarinids. Northward, at Site 432, the strongly impoverished foraminiferal assemblage (5 species) entirely consists of thermophilic species. Lower Eocene sediments from the middle part of the ridge (Site 433) yielded the foraminiferal assemblage of 18 species; the majority of them (72%) belong to the *Morozovella* and *Acarinina* genera, whereas others are represented by low-conical *Subbotina* and *Planorotalites* forms. In the Obruchev Rise, the foraminiferal assemblage (21 species) is mainly composed of *Subbotina* and *Planorotalites* representatives (up 83% in total). The early Eocene assemblage of the Koryak–Kamchatka region (26 species), similarly to the late Paleocene one, slightly differs from those of lower-latitude areas. Thermophilic species of the genera *Morozovella* and *Acarinina* together represent 53% of the assemblage; remaining species belong mostly to subbotinids.

The middle Eocene assemblage of planktonic foraminifers is recorded in sediments of the Mid-Pacific Mountains (30 species), Schatsky and Obruchev rises (14 and 26 species, respectively), and of the Koryak–Kamchatka region (11 species). As is seen, the species diversity of foraminifers in the North Pacific region is

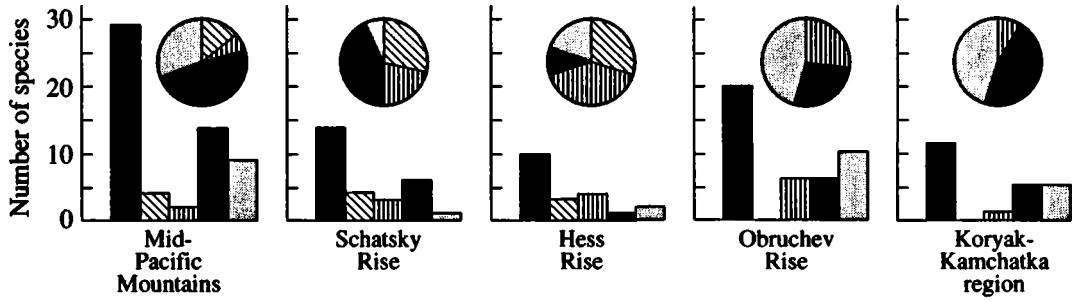

Fig. 5. Changes in the species diversity and composition of middle Eocene planktonic foraminifers in the North Pacific region (symbols as in Fig. 2).

less changeable as compared with the previous periods (Fig. 5). Simultaneously, the middle Eocene assemblage differs in its structure from that of the early Eocene time. It is composed of both the typical thermophilic (*Morozovella* and *Acarinina* genera) and the known-to-be cold-resistant (*Globorotaloides* and *Catapsydrax* genera) forms. Along with the above, the middle Eocene assemblage is marked by the appearance of moderately thermophilic species characteristic of subtropical conditions, such as the representatives of the *Pseudohastigerina* genus, *Turborotalia cerroazulensis* group, and high-conical *Subbotina* species, which are present in a subordinate amount in the middle Eocene assemblages of the the entire North Pacific region.

In the Mid-Pacific Mountains and Schatsky Rise, the middle Eocene planktonic foraminiferal assemblages (29 and 14 species, respectively) include mainly species of the *Morozovella*, *Acarinina*, *Globigerinatheka*, and other thermophilic (tropical) genera. They compose 55 and 79% of the assemblages, respectively, whereas the share of cold-resistant species does not exceed 31 and 14%. The temperate species play a subordinate role in these two areas (14 and 7%, respectively). In the southern part of the Hess Rise (Site 466), the foraminiferal assemblage consists of 10 species, seven of which (70%) belong to the *Morozovella* and *Acarinina* genera (i.e., are thermophilic), one species (10%) is temperate, and the remaining two species (20%) are cold-resistant. The strongly dissolved and, thus, impoverished middle Eocene (probably, latest early Eocene) assemblage recorded in the northern extremity of the Hess Rise (Site 310) is represented by five species, two of which (40%) belong to the *Acarinina* genus, one species is temperate, and two others, cold-resistant. In the assemblage of 22 species from the high-latitude North Pacific (the Obruchev Rise), the share of tropical elements decreases to 30%. Almost half of this assemblage (45%) is represented by the cold-resistant species. The remaining 25% of the assemblage are the temperate species. The assemblage of planktonic foraminifers (11 species) that is characteristic of cooler waters is discovered from the middle Eocene deposits of the Koryak–Kamchatka region. There, three quarters of the assemblage consist of typical cold-resistant *Subbotina* and *Globorotaloides* species, and other forms represent the temperate (subtropical) species.

Thus, the middle Eocene time in the North Pacific was marked by the development of latitudinal (climatic) differentiation of planktonic foraminiferal assemblages. This phenomenon commenced probably as early as in the early Eocene, most likely in the second half of this unit, when the temperature of surficial waters gradually decreased after the climatic optimum. This is evident, in particular, from the fact that, in high latitudes (the Koryak–Kamchatka region), where planktonic species are particularly susceptible to climatic fluctuations, the sharpest changes in foraminiferal assemblages occurred in the early Ipresian (Ben'yamovskii *et al.*, 1992; *Opornyi razrez...*, 1994).

Based on changes in the composition of planktonic foraminiferal assemblages along the south-north profile, the following biogeographic provinces can be distinguished in the North Pacific region for the middle Eocene time: tropical (the Mid-Pacific Mountains, Schatsky Rise, southern Hess Rise), subtropical (which is recognized conventionally because of insufficient data and comprises probably the northern part of the Hess Rise and most of the Emperor Seamounts), transitional (Obruchev Rise), and boreal (Koryak–Kamchatka region) provinces.

In the late Eocene, changes in the foraminiferal fauna become even more distinct. This is reflected both in a substantial diversity decrease at all sites and in the

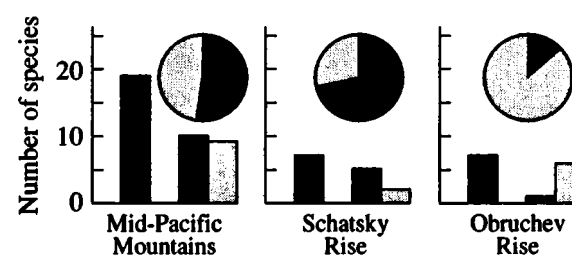

Fig. 6. Changes in the species diversity and composition of late Eocene planktonic foraminifers in the North Pacific region (symbols as in Fig. 2).

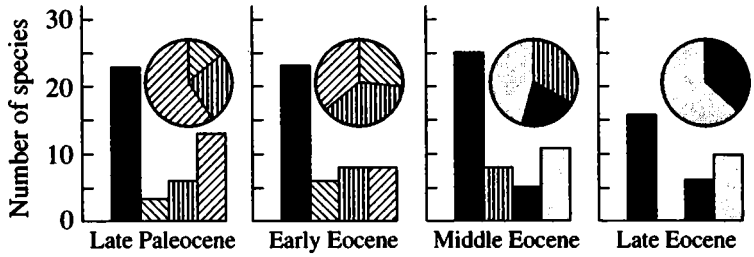

Fig. 7. Changes in the species diversity and composition of planktonic foraminifers in the Campbell Plateau area during the late Paleocene–Eocene (symbols as in Fig. 2).

structure of foraminiferal assemblages. In the North Pacific region, late Eocene foraminifers are recorded in the Mid-Pacific Mountains, Schatsky and Obruchev rises, as well as in the Koryak–Kamchatka area. They are most diverse (19 species) in the first region; in others, the number of species varies from 7 in the south to 4 in the north (Fig. 6). The thermophilic assemblage is discovered only in the Mid-Pacific Mountains, where more than a half of the assemblage is represented by thermophilic (53%) and temperate forms. In the Obruchev Rise area, the foraminiferal assemblage mainly consists of representatives of the genera *Globorotaloides*, *Catapsydrax*, *Globigerina*, and *Tenuitella* accompanied by rare specimens of temperate *Dentoglobigerina galavisi*. Single specimens of planktonic foraminifers discovered from the upper Eocene deposits of the Koryak–Kamchatka region (Komandor Islands) belong to the cold-resistant *Tenuitella* genus.

Thus, the late Paleocene–early Eocene planktonic foraminiferal assemblages are distinctly thermophilic throughout the entire North Pacific. The coeval fauna of the Koryak–Kamchatka region demonstrates the subtropical affinity. The latitudinal (climatic) differentiation becomes well-manifested in the middle Eocene, when the North Pacific region (northern Kamchatka included) was populated by climatically different assemblages: by the tropical (Mid-Pacific Mountains, Schatsky Rise, southern Hess Rise), subtropical (the northern part of the Hess Rise), transitional (Obruchev Rise), and boreal (Koryak–Kamchatka region) assemblages.

According to plate-tectonic reconstructions, the Obruchev Rise together with the Pacific lithosphere plate moved during the entire Early Cenozoic from 30° to 40°N, i.e., within the recent subtropical realm (Rea *et al.*, 1995). Nonetheless, the late Paleocene–early Eocene foraminiferal fauna reveals the tropical affinity. The subtropical assemblage is discovered only in the Koryak–Kamchatka region. This may imply the past biogeographic zoning quite different from that observable now. It seems likely that no boreal province existed in the North Pacific, as is the case with the natal province in the southern hemisphere. According to the oxygen-isotope analysis of planktonic foraminifers from the Mod Rise in the Antarctic region (Stott *et al.*, 1990), the surface temperature in the late Paleocene–early Eocene ocean was up to 18°C. Such a temperature in the modern ocean is characteristic of the subtropical realm in both hemispheres.

Diagrams plotted for the late Paleocene–Eocene foraminiferal assemblages of the Campbell Plateau (Site 275, modern latitude 50°S) illustrate the situation similar to that in the North Pacific (Fig. 7). In this area, which was located near the Antarctic in the Early Cenozoic time, the planktonic foraminiferal diversity experienced no substantial variations during the Late Paleocene–early Eocene (Jenkins, 1971; Krasheninnikov and Basov, 1986; Krasheninnikov *et al.*, 1988). More significant changes occurred in the structure of the assemblages. The late Paleocene and early Eocene assemblages in this area, similarly to those from the high northern latitudes, reveal the thermophilic affinity (subtropical in the late Paleocene and tropical in the early Eocene). Species of the genera *Morozovella* and *Acarinina* represent 41 and 63% of the assemblages, respectively, in these time intervals. The other part was mainly represented by *Subbotina* and *Chiloguembelina* forms.

Biogeographic differentiation of planktonic foraminifers in the southern hemisphere commenced, similarly to the northern one, in the middle Eocene and was particularly distinct in the late Eocene. It is interesting that planktonic foraminifers from both the Campbell Plateau and Obruchev Rise reveal the subtropical affinity, though the share of thermophilic species in the former area is slightly lower than in the latter one. However, it should be kept in mind that the Campbell Plateau was located at higher latitudes than nowadays, and, to the contrary, the Obruchev Rise occupied significantly lower latitudes within the modern subtropical realm (Rea *et al.*, 1995). Similar distribution patterns were also established in the Atlantic (Boersma and Premoli Silva, 1991). This means that the middle Eocene foraminiferal assemblages from the high latitudes of the southern hemisphere were slightly more thermophilic as compared with coeval assemblages from the northern hemisphere, i.e., asymmetry in distribution of present climatic belts appeared probably later in response to the initiation of the Circum-Antarctic Current.

CONCLUSION

The analysis of diversity, taxonomic composition, and structure of planktonic foraminiferal assemblages along with changes in these parameters within the Paleogene sections of the North Pacific region, from the Mid-Pacific Mountains in the south (19–21°N) to northern Kamchatka (60°N) in the north, revealed that, during the late Paleocene–early Eocene, the entire North Pacific was populated by a rather uniform foraminiferal assemblage dominated by representatives of thermophilic (tropical) *Morozovella* and *Acarinina* genera. Their dominating role was less pronounced only in the late Paleocene–early Eocene assemblage from the Il'pinskii Peninsula of northern Kamchatka, i.e., in the northernmost section. Differences in composition of planktonic foraminifers from different latitudes, against the background of their unchanged or somewhat higher diversity, became evident beginning from the middle Eocene, when groups of thermophilic (tropical), temperate (subtropical), and cold-resistant (boreal) species can be distinguished among planktonic foraminifers. The latitudinal differentiation and the climatic confinement of planktonic foraminiferal assemblages commenced apparently earlier, following the temperature decrease after the early Eocene climatic optimum, and became particularly distinct in the late Eocene.

A similar trend of changes in species diversity and composition of planktonic foraminiferal assemblages during the Paleocene and Eocene is characteristic of high latitudes of the southern hemisphere. The comparison of late Paleocene and early Eocene assemblages from the Obruchev Rise and Campbell Plateau located nowadays at similar latitudes in the northern and southern hemispheres, respectively, shows that in the latter area the preponderance of tropical species was slightly lesser than in the North Pacific.

ACKNOWLEDGMENTS

The work was supported by the Russian Foundation for Basic Research, project 96-05-66269.

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The Lower Karangat Palynostratigraphy in Abkhazia

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Received December 21, 1995; in final form November 12, 1996

Abstract—The detailed climatostratigraphic subdivision, which is first suggested for the lower Karangat deposits in Abkhazia, depicts four warming stages and three intermediate cooling periods in the climatic evolution of the region. This conclusion is based on results of palynological analysis of 6.5-m-thick sequence of the lower Karangat (Riss–Würm) sediments recovered by drilling near the town of Sukhumi from the depth interval of 38.5–45 m. It is established that distribution patterns of vegetation during the respective period of sedimentation have not been persistent. Formations of dark coniferous and broad-leaved plants, the latter confined to the middle and lower belts of the mountain vegetation, periodically migrated in the past. Dynamics of this migration corresponds to three stages of climatic evolution, and the last of them is divisible in four substages and several climatic phases.

Key words: *palynostratigraphy, Karangat time, climatic evolution, flora migration.*

INTRODUCTION

The lower Karangat deposits bearing molluscan fauna are widespread in the western part of Georgia (Dzvelaya, 1956; Mamaladze, 1975; Fedorov, 1978). The Karangat flora, represented by leaf impressions and first discovered in deposits of the Mt. Sukhumskaya and Lechkopi areas (Kolakovskii, 1952; Ratiani, 1959), was originally described as including 42 taxa (Fig. 1). Afterward, their list was supplemented with new leaf species identified in sediments of the Gumista area (Ratiani, 1960, 1970; Ratiani and Solov'ev, 1965) and in travertine beds of the Higher Svanetia (Ratiani, 1964).

Additional palynological data on the Karangat flora were obtained after investigation of six samples from the Pleistocene deposits of the Khorg, Pitsunda, and Kobuleti areas (Shatilova, 1974; Shatilova and Badzoshvili, 1966). Later, palynomorphs of the Karangat Horizon were studied in nine samples from the Patara Poti area (Mamatsashvili, 1975) and in sixteen samples from the Sukhumi area (Kvavadze *et al.*, 1984). In the last work, we arrived at the conclusion that the dominant type of vegetation of the Karangat period corresponded in general to warmer and less humid climatic conditions as compared to the present-day environments. At the same time, the high abundance of microphytofossils and pollen diversity in the studied samples stimulated us to investigate them in detail in order to elaborate the more sophisticated stratigraphic scheme of the lower Karangat in Abkhazia. This objective and the elucidation of some important stages in evolution of the early Karangat climate are the principal goals of our current study.

MATERIAL AND INVESTIGATION METHODS

Using the drilling rig URB-3AM (UKS-22M), we penetrated through deposits of the 1st Black Sea terrace of the East Sukhumi Cape down to the depth of 44.7 m (Well 36, 43°N and 57°49′E) obtaining core samples 108 mm in diameter. The total number of samples selected for palynological analysis is 120. The complete characteristics of the Holocene interval (upper 26 m) and data on its stratigraphic subdivision have already been published (Kvavadze *et al.*, 1984; Kvavadze and Rukhadze, 1989), so we will consider below only the pollen spectra from the lower part of the well. Sediments of this interval are represented by fine-grained sand with intercalations of gravel and pebbles. We consider the lowermost beds in the depth interval of 38.5–45.0 m as corresponding to the Riss–Würm Interglacial or the early Karangat period following the conclusions of Maruashvili (1985, 1991) and Zubakov (1986), who correlate the late Karangat with the Würm glacial epoch.

Our inference about both age intervals is based on the ^{14}C date of 6425 ± 60 years (TB-352), obtained for the sample from the depth of 8.5 m, and also on the character of spore–pollen spectra (SPS). Below the indicated boundary, the latter reveal a decline of formerly abundant Taxodiaceae, *Cedrus*, *Carya*, and *Tsuga* simultaneous to the growing abundance of *Carpinus, Castanea, Fagus, Quercus, Tilia*, and *Ulmus* pollen indicating the preponderance of broad-leaved forests. In contrast, spore–pollen assemblages (SPA) from higher beds are extremely rich in *Abies* pollen (up to 70%) implying the very low altitudinal position of dark coniferous forests and, subsequently, a considerable climatic cooling typical of the Würm glaciation.

We would also like to note that in our stratigraphic subdivision and discrimination of palynozones in mountain regions of Transcaucasia we rely primarily upon dominant taxa of SPS and, to some extent, on the tendency of growing or decreasing content of pollen representing plant groupings different in ecological aspect. For instance, pollen of *Abies* and *Fagus* is the most clear indicator of climatic fluctuations, because its growing abundance in deposits of the coastal zone suggest that forests of this type approach the sedimentation site and thus reflect a deterioration of climatic environments (Fig. 2). In such a case, pollen spectra become simultaneously depleted in pollen species of *Carpinus*, *Castanea*, *Quercus*, *Tilia*, and other broad-leaved genera, whereas the amelioration of the climate leads to the opposite distribution patterns of pollen taxa. In addition, we want to stress that the terms "cold" and "arid" are used in the text in their conventional, relative meaning. Other criteria of discriminating pollen assemblages known in Transcaucasia are described in our earlier publications (Kvavadze and Rukhadze, 1989; Kvavadze, 1993, 1994).

We studied 24 rock samples characterizing the 6.5-m-thick sediments of the lower Karangat interval; however, only 16 of them (from beds of fine-grained sand) yielded the sufficient amount of well-preserved microphytofossils. The preliminary treatment of samples was performed by using the alkaline and acetol maceration. The pollen diagram in Fig. 3 is plotted by using the computer program "Tilia-Graph" of E. Grimm. Percentages of individual components of pollen spectra are calculated relative to the total content of pollen and spores taken for 100%. Another diagram (Fig. 3) is plotted separately for the arboreal pollen only.

INVESTIGATION RESULTS

The spore–pollen diagram obtained (Fig. 3) shows the preponderance of arboreal plants and the low content of spores throughout the studied sedimentary sequence, along with a rather high abundance of herbaceous pollen in its lower and upper parts. According to peculiar features of palynomorph distribution at different depth levels, we distinguished 10 SPA, whose characteristics and depth intervals (in parentheses) are as follows.

SPA-1 (45.0–44.4 m). 60% of arboreal pollen, 34% of herbaceous taxa, and 6% of spores. The dominant *Carpinus* pollen (27–31%) coexists with rather abundant specimens of *Pinus* and *Alnus* (12–23% of each taxon). The summary content of broad-leaved taxa is 36–42%. The abundance of *Abies* equals 5–13%, and that of *Picea*, 1%. Exotic Taxodiaceae, *Carya*, *Tsuga*, and *Cedrus* represent up to 10%. Herbaceous group includes Compositae (39%), Gramineae (29%), and *Artemisia* (10%).

SPA-2 (44.4–43.8 m). Arboreal pollen is less abundant (52%), but the content of herbaceous taxa is higher

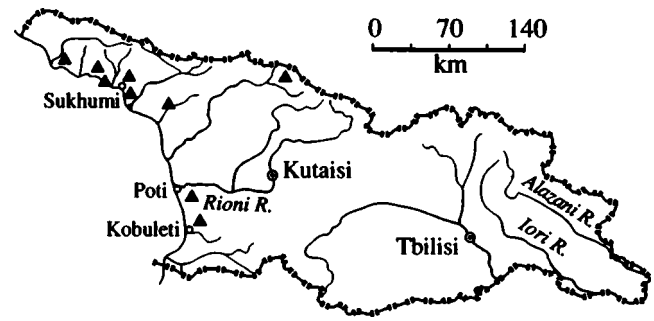

Fig. 1. Map of Georgia with localities of the studied Karangat sections.

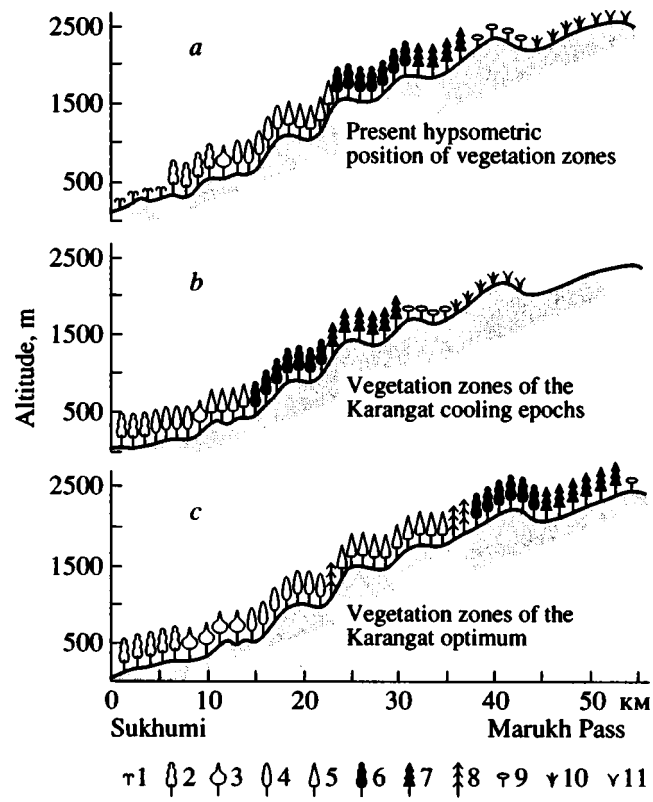

Fig. 2. Position of vegetation zones in the Sukhumi–Marukh Pass transect: (*a*) Recent epoch; (*b*) Karangat cooling phases; (*c*) Karangat warming phases; (1) psammophilic vegetation; (2) alder; (3) oak; (4) chestnut; (5) hornbeam; (6) beech; (7) fir; (8) pine; (9) Subalpine assemblages; (10) Alpine assemblages; (11) high-mountain tundra plants.

(43%). *Alnus* (24%) dominant at the beginning yields its abundance rank afterward to *Abies* (30%). Percentages of *Pinus* (8–19%) and broad-leaved taxa in total (29–32%) are decreased, although *Carpinus* (10–19%) is still dominant among the latter. *Picea* and *Betula* constitute 2% each, but pollen of exotic taxa (4%) is less abundant. Compositae (52%) turn out to be more copious among herbs in contrast to lower content of *Artemisia* (4%); this group also includes Cyperaceae (3%).

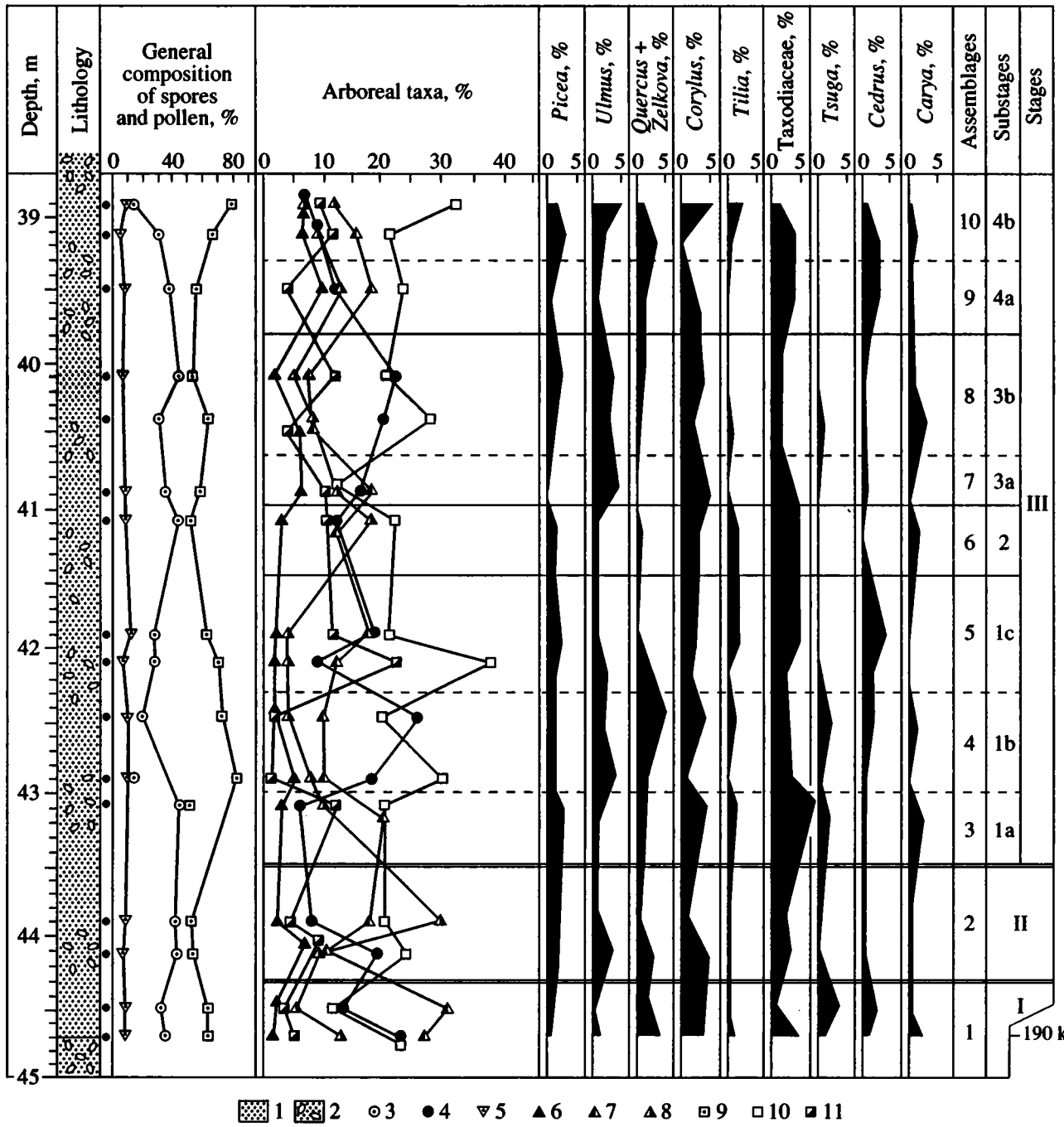

Fig. 3. Spore–pollen diagram for arboreal and fruticose taxa from the lower Karangat deposits of Sukhumi area: (1) sand; (2) sand with gravel and pebbles; (3) herbs; (4) *Pinus*; (5) spores; (6) *Fagus*; (7) *Carpinus*; (8) *Larix*; (9) arboreal taxa; (10) *Alnus*; (11) *Castanea*.

SPA-3 (43.8–43.0 m). Proportions of taxa in the general pollen composition are almost unchanged. The total content of broad-leaved taxa, among which *Carpinus* (20%) is dominant and *Castanea* (11%) subdominant, increases up to 40%. The abundance of *Abies* (10%) significantly decreases; the percentage of *Picea* pollen is 3%. Exotic taxa, excluding *Cedrus* (1%), are most abundant (13%). Compositae pollen represents 61%, and Chenopodiaceae fraction amounts to 16%.

SPA-4 (43.0–42.3 m). 73–82% of arboreal pollen, 10–18% of herbaceous taxa, and 8–9% of spores. Pollen of *Alnus* (20–30%), and then of *Pinus* (18–26%) is dominant. *Carpinus* (10%) prevail in the group of broad-leaved taxa (23% in total). *Quercus*, *Zelkova*,

and *Ulmus* turn out to be more abundant than before, whereas the content of exotic pollen decreases. Gramineae (50%) represent the main fraction of herbaceous pollen.

SPA-5 (42.3–41.7 m). 62–69% of arboreal specimens, 27% of herbaceous taxa, and 4–11% of spores. *Alnus* (21–38%) is dominant. Castaneae (2–12%), first, and, then, *Carpinus* (12–17%) preponderate amid broad-leaved forms (41% in total). The pollen content of *Pinus* (9–17%) and *Abies* (4%) diminished, but the abundance of Taxodiaceae (up to 5%) and *Cedrus* (up to 4%) became higher. Beginning from this level and higher in the section, pollen of *Tsuga* is missing (except several grains in SPA-7). Herbaceous pollen is more diverse.

SPA-6 (41.7–41.0 m). 51% of arboreal taxa, 43% of herbaceous pollen, and 4% of spores. *Alnus* (22%) and *Abies* (18%) represent dominant genera. *Pinus*, *Carpinus*, and *Castanea* occur in equal proportions (12% each). Broad-leaved forms constitute 32% in total. Compositae are very abundant (79%).

SPA-7 (41.0–40.7 m). 58% of arboreal pollen, 34% of herbaceous specimens, and 8% of spores. *Carpinus* (17%) and *Pinus* (16%) are dominant forms associated with less abundant *Alnus*, *Abies*, and *Castanea* (12% each). Broad-leaved taxa represent 40% of the assemblage. Taxodiaceae (4%) are most abundant among exotic specimens. Gramineae (22%) and Chenopodiaceae (13%) are more plentiful than before in contrast to less numerous Compositae (59%).

SPA-8 (40.7–40.0 m). 50–62% of arboreal pollen, 31–44% of herbaceous species, and 7% of spores. *Alnus* (28% at the beginning and 21% after) associates with *Pinus* (22%) in the dominant group. The abundance of *Carpinus*, *Castanea*, *Fagus*, and *Ulmus* decreases, and, together with other broad-leaved taxa, they constitute 22–27% of the assemblage. Additional components are represented by *Abies* (5–8%), *Picea* (3%), and exotic palynomorphs (6%). Pollen of Compositae further decrease in abundance (down to 47%), but Gramineae specimens are more opulent (up to 30%).

SPA-9 (40.0–39.4 m) is almost the same in general composition. *Alnus* (23%) is again dominant. Broad-leaved taxa (37% in total) are mainly represented by *Carpinus* (18%), *Fagus* (11%), and *Castanea* (4%). They coexist with *Abies* (13%), *Pinus* (12%), and exotic forms (8%). The herbaceous group once again includes Cyperaceae (3%).

SPA-10 (39.4–38.7 m). 65–79% of arboreal pollen, 12–31% of herbaceous species, and 4–9% of spores. The assemblage includes *Alnus* (21–32%), *Carpinus* (11–15%), *Castanea* (9–11%), *Fagus* (6–7%), and other broad-leaved taxa (34–39% in total) in association with *Abies* (7–9%), *Pinus* (7–9%), *Picea* (2–3%), and exotic species (3%). Compositae (59%) and Gramineae (19%) are main representatives of herbs.

Thus, the predominance of arboreal pollen (50–82%) is characteristic of all the assemblages, which are also rather rich (10–45%) in herbaceous specimens (Fig. 4). *Achillea*, *Salsola*, and *Artemisia* dominant among the latter are indicators of psammophilic vegetation typical of seaside sands (Fig. 2). The scarcity of sedge and buckwheat group, along with relatively low abundant (as compared to Holocene) alder pollen, suggests that lowland marshes of the present-day type were absent in the past. *Juncus*, *Typha*, and *Sparganium* represent taxa that formerly populated lagoon lakes. Spores are not abundant and mainly belong to Polypodiaceae and *Pteris cretica*. Pine pollen is quite significant in the composition of the assemblages. The parallel decrease in the abundance of broad-leaved pollen in SPA-4 and SPA-8 may indicate that the site of sedimentation was remote from broad-leaved forests during the expansion stage of pine and alder communities. The variable character of distribution of arboreal pollen seems to be evidence of inconstant forest types during the early Karangat time in response to secular climatic fluctuations. The section reveals three horizons with a rather high content of *Abies* pollen (30, 18, and 13%), which means a deterioration of climatic conditions. All the changes in vegetation composition and climatic environments reflect three stages in their evolution, and the last stage is divisible into four substages and several climatic phases, as is illustrated in the suggested scheme of detailed climatostratigraphic subdivisions of the Riss–Würm in Abkhazia (Fig. 3).

Stage I (SPA-1). Dominant hornbeam forests of the stage included accessory taxa, initially, *Castanea*, and, later, parallel to the humidity increase, *Castanea* and *Fagus* forms, which became equally abundant. In the plain areas, *Carpinus* coexisted with *Alnus*, but at the end of the stage the first taxon turned out to be dominant even here, and alder, together with *Pterocarya* admixture, presumably populated the river floodplains. In accordance with humidity variation, pine forests of the initial time were more widespread than later. In addition, the vegetation cover included *Zelkova*, *Quercus*, Taxodiaceae, *Carya*, *Tsuga*, and *Cedrus*. Dark coniferous forests migrated upward. The moderately warm and not so humid conditions changed with time to more warm and humid.

Stage II (SPA-2). Alder and pine forests became dominant. Humidity decreased. The hornbeam population considerably reduced, while *Ulmus* appeared as an element of the vegetation cover. The former was a component of *Carpinus–Castanea–Fagus* forests, which migrated downward in response to invasion of dark conifers. The stage ended with the cooling phase of a higher humidity and lower position of the fir-forest zone, when the alder–hornbeam forests returned to lowlands, and *Castanea* and *Fagus* forms became rare in the broad-leaved forests. Areas of broad-leaved, pine, and Taxodiaceae forests reduced, and other exotic species were on the verge of extinction. The climate was relatively cold and humid.

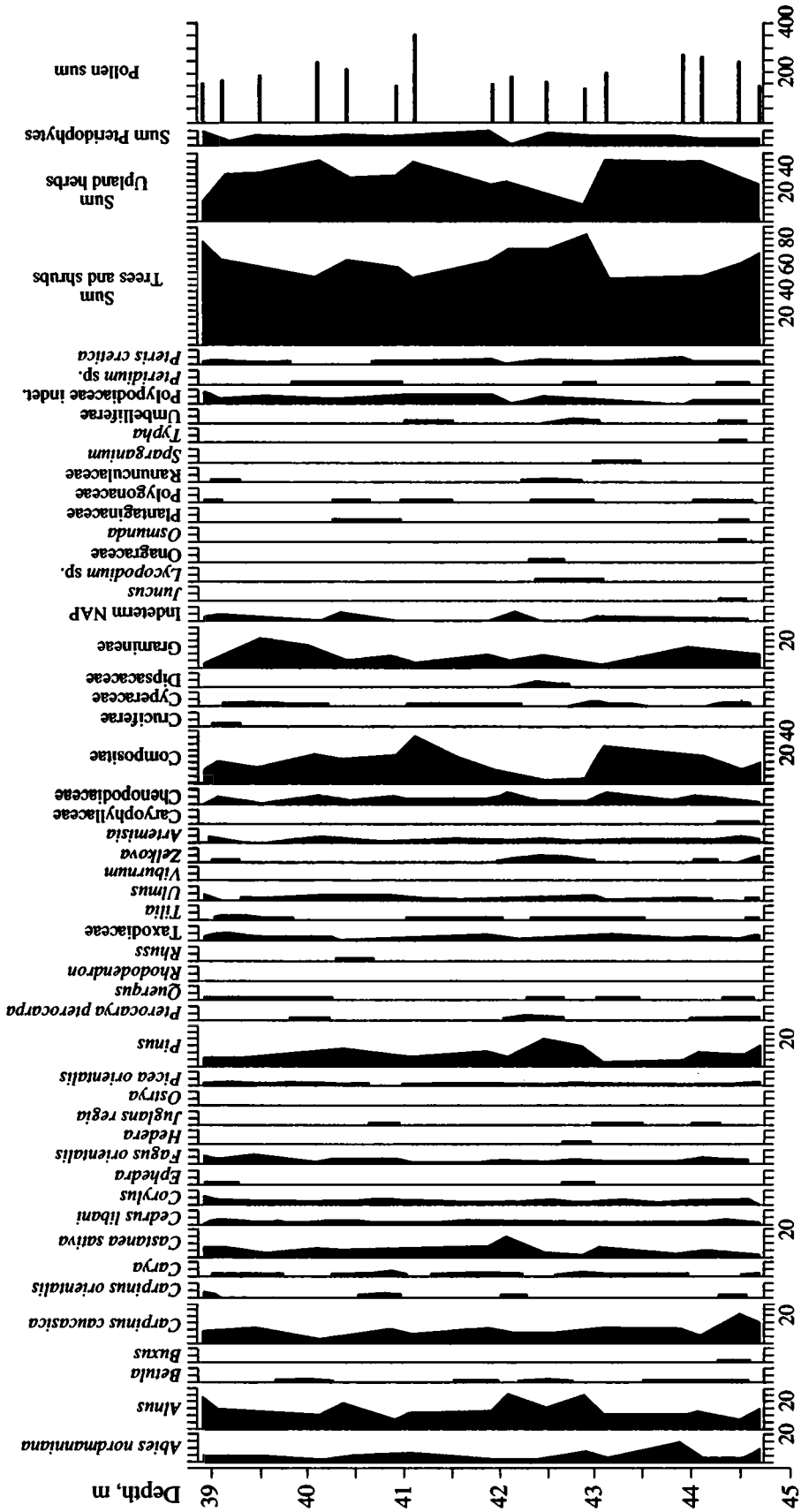

Fig. 4. Generalized spore–pollen diagram for the Karangat deposits of Sukhumi area.

Stage III (SPA-3 to SPA-10). This stage corresponds to the Riss–Würm climatic optimum divisible into four substages with complicated changes in climatic environments.

Substage 1 (SPA-3–5) represented a rather long period of warm climate. According to changes in vegetation distribution, we distinguish at least three climatic phases in this substage.

Phase 1a. Alder–hornbeam forests dominated in lowlands on soils of the gley Podsol type like their recent analogues (Dolukhanov, 1989). Areas of hornbeam forests with rather widespread *Castanea* forms grew and displaced upward the dark conifers. Taxodiaceae were at their abundance maximum. *Carya* and *Tsuga* became more abundant. The climate was moderately warm and humid.

Phase 1b. Pine forests were dominant, although alder thickets were more abundant at the beginning. Hornbeam forests with an insignificant admixture of *Castanea* and *Fagus* species migrated at a higher hypsometric level. The abundance of *Quercus*, *Ulmus*, *Zelkova*, and *Cedrus* grew, whereas Taxodiaceae and *Carya* became somewhat less widespread. The climate was very warm, and aridity was growing up to the maximum.

Phase 1c. A sudden humidity increase stimulated expansion of chestnut forests with an admixture of hornbeam trees. The maximum expansion of alder thickets decreased the distribution areas of pine forests. At the end of the phase, when environments became less humid, alder-populated areas became smaller, and *Carpinus* was more abundant than *Castanea*. A somewhat greater abundance than before was characteristic of Taxodiaceae and *Cedrus*, whereas the abundance of *Carya* was very low, and *Tsuga* became extinct. The climate was warm and sufficiently humid in general.

Throughout the substage, the dark coniferous forests were situated at a high altitude in the limited mountain areas.

Substage 2 (SPA-6). The downward migration of dark coniferous forest displaced the hornbeam–chestnut assemblages with *Fagus* and *Tilia* admixture into a lower zone. Alder forests stayed in the former status, whereas the pine-forest areas were lessening. Exotic forms were represented by Taxodiaceae and *Carya*. The climate was moderately cold and humid.

Substage 3 (SPA-7 and SPA-8) marking a period of new warming and aridity growth is subdivided into two phases.

Phase 3a. Hornbeam forests with quite abundant *Castanea* forms were dominant. The abundance of beech and elm trees grew. Areas of alder population lessened. Vegetation zones migrated upward, and areas of pine forests were expanding from this moment up to the end of the substage. The climate was moderately warm and humid.

Phase 3b. Pine forests and alder thickets again were dominant. Broad-leaved forests migrated upward and predominantly consisted of hornbeam trees with subordinate *Fagus* and *Castanea* forms. The last taxon became dominant at the end of the phase. *Quercus*, *Ulmus*, and *Carya* played a more significant role in the vegetation cover. The abundance of Taxodiaceae decreased. Spruce trees became more abundant in the dark coniferous forests. The climate was very warm, though not very humid. The phase marks the second peak of warming and aridization, presumably less significant than the first one.

Substage 4 (SPA-9 and SPA-10) characterizes the temperature decrease and parallel increase of humidity. It is also divisible into two phases.

Phase 4a. Vegetation zones migrated downward, and alder–hornbeam forests returned once more in plain areas. The population of *Fagus* trees in the hornbeam forests increased, while *Castanea* forms represented an insignificant admixture in this community. *Fagus* species presumably appeared in the dark coniferous forests, whose zone migrated downward. Exotic forms were rare. The climate was cool and humid.

Phase 4b. In the course of climatic warming and humidification, *Fagus* grouping in the hornbeam forests was giving its place to *Castanea* species. The phase terminated with expansion of hornbeam–chestnut forests containing an admixture of beech trees. *Ulmus* forms were also growing in abundance, and alder thickets became very widespread. Areas of pine forests were limited. Exotic taxa gradually disappeared. The climate was warm and more humid than before.

DISCUSSION

Using our and other data available at present (Table 2), we may conclude that hornbeam–alder forests with local alder thickets dominated in lowland landscapes of the region throughout the Riss–Würm or early Karangat period. According to macroremains, alder taxa were represented by *Alnus hoernesi* Stur and, to a lesser extent, by *A. barbata* C.A.M. *Pterocarya* of a limited abundance populated river valleys in association with the more abundant *Salix alba* L. Pine forests occupied important areas in the vegetation cover. Groupings of Taxodiaceae, *Cedrus*, *Carya*, and *Tsuga* were under gradual decline. Judging from macroremains, *Parrotia pristina* Ett. episodically was one of the taxa dominant in lowlands. *Zelkova* groupings similar in ecological features to *Quercus* assemblages sporadically populated some areas in the seashore plain and nearby piedmont slopes. A higher zone of drier and better illuminated mountain slopes was occupied by oak forests (*Quercus cerris* L.) with an undergrowth consisting of *Hypericum*, *Azalea*, *Carpinus*, *Lonicera*, *Pyracanta*, and *Crataegus*. Chestnut forests existed in shaded slopes and sometimes descended down to the seashore plain. *Fagus* occurred as an episodic admixture in the

Table 1. Scheme of the Riss–Würm climatostratigraphic subdivisions in western Georgia correlated with data of other researchers

Stage	Substage	Kvavadze and Rukhadze, 1989 (Sukhumi)	Mamatsashvili, 1975 (Potara Poti)	Kolakovskii, 1952 (Mt. Sukhumskaya, Lechkopi)	Ratiani, 1974 (Gumista River)	Shatilova and Badzoshvili, 1966 (Kobuleti)	Climate
III Stage. Riss–Würm climatic optimum	4b	Hornbeam–chestnut forests, alder thickets			*Abies cephalonica* thickets and *Carpinus–Fagus–Castanea* forests, thickets of *Alnus hoernesi* and less abundant *A. barbata*, forests of *Parrotia pristina*, reduction of xerophilous elements down to three species		Warm and humid
	4a	Migration of vegetation zone downward, hornbeam forests with beech admixture, alder–hornbeam forests in plain areas	Beds with maximum abundance of beech and hornbeam pollen, alder pollen is also rather abundant				Cool and humid
	3b	Predominance of pine and alder forests, hornbeam forests with admixture of chestnut trees	Beds with maximum abundance of *Carpinus* pollen, quite abundant pollen of pine and chestnut, content of alder pollen is decreasing			Pine and oak–chestnut forests, assemblages with predominant *Fagus* and subordinate *Tilia* and *Acer* species, alder thickets	Warm, somewhat less humid
	3a	Hornbeam forests with abundant chestnut admixture, alder thickets	Beds with maximum abundance of chestnut, alder, and linden pollen				Moderately warm and humid
	2	Migration of dark coniferous forests downward, hornbeam–chestnut forests with admixture of beech trees, alder thickets	Beds with maximum abundance of beech pollen and high content of alder pollen				Moderately cold and humid
	1c	Chestnut forests at the beginning and then hornbeam forests with admixture of chestnut and local alder thickets	Beds with predominant pollen of hornbeam, beech, chestnut, hazel, and elm				Warm and humid
	1b	Pine and hornbeam forests, growing significance of oak and elm	Beds with maximum abundance of pine and spruce pollen, abundance of mesophilic taxa (*Fagus*, *Tilia*, and others) is decreased	*Quercus cerris* forests, formations of the Mediterranean maquis type (14 species), forests of shaded canyons with chestnut, beech, maple, box trees, ivy, and other plants			Very warm with growing aridity
	1a	Alder–hornbeam and hornbeam forests, the latter with admixture of chestnut trees					Moderately warm and humid

Table 2. Taxonomic composition of the early Karangat flora

Plants	Data from Kolakovskii (1952)	Data from Ratiani (1964)	Data from Shatilova and Badzoshvili (1966)	Data from Mamatsashvili (1975)	Data from Kvavadze (1984)	Plants	Data from Kolakovskii (1952)	Data from Ratiani (1964)	Data from Shatilova and Badzoshvili (1966)	Data from Mamatsashvili (1975)	Data from Kvavadze (1984)
1	2	3	4	5	6	1	2	3	4	5	6
Type Sphagnum	–	–	–	+	–	*Juglans regia* L.	–	–	+	+	+
Type Bryales	–	–	–	+	–	*Carya aquatica* (Michx.) Nutt	–	–	+	–	+
Licopodium clavatum L.	–	–	+	+	+	*Salix alba* L.	–	+	–	–	–
L. selago L.	–	–	–	–	–	*S. caprea* L.	–	+	–	–	–
Selaginella selaginoides (L.) Link.	–	–	+	+	–	*S. daphnoides* Vill.*	+	–	–	–	–
Bothrychium sp.	–	–	+	–	+	*S. micans* Anderss.*	+	–	–	–	–
Ophioglossum sp.	–	–	+	–	–	*S. triandra* L.	+	–	–	+	–
Osmunda regalis L.	–	–	+	+	+	*Salix* sp.	–	–	–	+	–
Cryptogramma crispa (L)R. Br.	–	–	+	–	+	*Populus alba* L.	+	+	–	–	–
Pteris cretica L.	–	–	+	+	+	*P. balsamoides* Gopp.*	+	+	–	–	–
Asplenium trichomanes L.	–	–	–	+	–	*P. tremula* L.	+	+	–	–	–
Polypodium australis Fee	–	–	–	–	–	*Alnus barbata* CAM	+	+	–	+	+
P. serratum (Willd.) Sauter	–	–	–	+	–	*A. glutinosa* (L.) Gaertn.	–	–	+	+	–
P. vulgare L.	–	–	+	–	–	*A. hoernesi* Stur*	–	+	–	–	–
Polypodium sp.	–	–	+	–	–	*Betula litwinowii* A. Dol.	–	+	–	–	–
Polypodiaceae gen. indet.	–	–	–	+	–	*Betula* sp.	–	–	–	+	+
Dryopteris filix-mas (L.) Schott	–	–	–	+	–	*Carpinus caucasica* A. Grossh.	–	–	+	+	+
D. oreopteris (Ehrh.) Maxon	+	–	–	–	–	*C. betulus* L.	+	+	–	–	–
Dryopteris sp.	+	–	–	–	–	*C. orientalis* Mill.	+	+	+	+	–
Pteridium sp.	–	–	–	+	–	*Corylus avellana* L.	+	–	+	–	–
Thelypteris sp.	–	–	+	–	–	*Corylus* sp.	–	–	–	+	+
Abies cephalonica Loud*	–	+	–	–	–	*Ostrya carpinifolia* Scop.	–	+	–	–	–
A. nordmanniana (Stev.) Spach.	–	+	+	+	+	*Fagus antipovii* Heer*	–	+	–	–	–
Cedrus deodara Loud*	–	–	+	–	+	*F. orientalis* Lipsky	+	+	+	+	+
Picea orientalis L.	–	–	+	+	+	*Castanea atavia* Ung.*	–	+	–	–	–
Pinus sosnowskii Nakai	–	+	–	–	–	*C. sativa* Mill.	+	–	+	+	+
Pinus sp.	–	–	+	+	+	*Quercus* aff. *iberica* Stev.	–	–	+	–	–
Tsuga europea (Menzl.) Szafer*	+	–	–	–	–	*Q.* aff. *pontica* C. Koch.	–	–	+	–	–
Tsuga sp.*	–	–	–	–	+	*Q. pseudorobur* Kov. (*Q. roburoides* Bereng)*	+	+	–	–	–
Thuja occidentalis L.*	+	–	–	–	–	*Q. cerris* L.	+	+	–	–	–
Taxodium sp.*	–	–	–	+	–	*Q.* cf. *ilex* L.	+	–	–	–	–
Taxodiaceae gen. indet.*	–	–	+	–	+	*Quercus* sp.	–	–	–	+	+
Ephedra sp.*	–	–	+	–	+	*Ulmus foliaceae* Gilib.*	+	+	+	–	–
Pterocarya pterocarpa (Michx.) Kunth	+	+	+	+	+	*U. scabra* Mill.*	–	–	+	–	–

Table 2. (Contd.)

1	2	3	4	5	6	1	2	3	4	5	6
Ulmus sp.	–	–	–	+	+	*Buxus colchica* A. Pojark.	+	–	–	+	–
Zelkova carpinifolia (Pall.) Dipp.*	+	+	+	+	+	*B. sempervirens* L.	–	+	–	–	–
Morus alba L.*	–	–	+	–	–	*Rhamnus* sp.	–	–	+	–	–
Ficus sp.	–	–	–	+	–	*Vitis* sp.*	+	–	–	–	–
Moraceae gen. indet.	–	–	–	+	–	*Tilia caucasica* Rupr.	–	+	+	+	+
Urtica sp.	–	–	–	+	–	*T. ledebourii* Borb.	–	–	+	–	–
Polygonum persicaria L.	–	–	+	–	–	*Daphne* sp.	–	–	+	+	–
P. viviparum L.	–	–	+	–	–	*Elaeagnus* sp.*	–	–	+	–	–
Polygonum sp.	–	–	–	+	+	*Viola* sp.	–	–	–	+	–
Caryophyllaceae gen. indet.	–	–	+	+	+	*Epilobium* sp.	–	–	+	–	–
Salsola sp.	–	–	–	–	+	*Trapa* cf. *colchica* N. Alb.	–	–	+	–	–
Chenopodiaceae gen. indet.	–	–	+	+	+	*Cornus* sp.	–	–	+	+	–
Magnolia geogrica Kol.*	+	–	–	–	–	*Hedera colchica* C. Koch.	+	+	–	–	+
Ranunculaceae gen. indet.	–	–	–	+	–	*H. helix* L.	+	–	–	–	–
Laurus nobilis L.*	+	+	–	–	–	*Turgenia latifolia* (L.) Hoffm.	–	–	+	–	–
Nuphar luteum L.	–	–	+	+	–	Umbelliferae gen. indet.	–	–	+	+	–
Nympheaceae gen. indet.	–	–	–	+	–	*Rhododendron flavum* Don.	+	–	–	–	–
Papaveraceae gen. indet.	–	–	–	+	–	*Rh. ponticum* L.	+	–	–	–	–
Cruciferae gen. indet.	+	–	–	–	–	*Rhododendron* sp.	–	–	+	+	+
Hypericum. inodorum Willd.	–	–	–	+	–	*Arbutus andrachne* L.	+	–	–	–	–
Platanus sp.*	–	–	–	–	+	*A. unedo* L.*	+	+	–	–	–
Parrotia pristina Ett.*	–	+	–	–	–	Ericaceae gen. indet.	–	–	–	+	–
Rosa sp.	–	–	+	+	–	*Diospyros brachisepala* A. Br.*	+	–	–	–	–
Crataegus sp.*	+	–	–	–	–	*Styrax officinalis* L.	–	+	–	–	–
Pyracantha coccinea Roem.	+	+	–	–	–	*Periplica graeca* L.	+	–	–	–	–
Sanguisorba sp.	–	–	+	–	–	*Fraxinus* sp.	–	–	–	+	–
Sorbus aucuparia L.	–	+	–	–	–	*Ligustrum* sp.	–	–	–	–	+
S. subfusca (Ledeb.) Boiss.	–	+	–	–	–	*Lonicera caprifolium* L.	+	–	–	–	–
Sorbus sp.	–	–	–	+	–	*Lonicera* sp.	–	–	–	–	+
Leguminosae gen. indet.	–	–	–	+	–	Labiatae gen. indet.	–	–	+	+	–
Geranium sp.	–	–	+	–	–	*Plantago* sp.	–	–	+	+	–
Mallotus cf. *apelta* Muller*	–	+	–	–	–	*Valeriana* sp.	–	–	+	–	–
Cotinus coggigria Scop.	+	–	–	–	–	*Cephalaria* sp.	–	–	+	–	–
Rhus coriaria L.	+	–	–	–	–	*Knautia* sp.	–	–	+	–	–
Rhus sp.	–	–	+	+	+	Dipsacaceae gen. indet.	–	–	–	+	–
Acer laetum C.A.M. *pliocenicum**	–	+	–	–	–	Campanulaceae gen. indet.	–	–	–	+	–
A. ibericum M.B*	+	–	–	–	–	*Artemisia* sp.	–	–	+	+	+
A pseudoplatanus L.	+	+	–	–	–	*Achillea* sp.	–	–	–	–	+
A. aff. *platanoides* L.	–	–	+	–	–	Compositae	–	–	+	+	+
A. polymorphum pliocenicum Sap.*	+	+	–	–	–	*Smilax excelsa* L.	–	+	–	–	–
A. productum A.Br.*	–	+	–	–	–	*Fritillaria* sp.	–	–	–	+	–
A. trautvetteri Medw.	–	+	–	–	–	Liliaceae gen. indet.	–	–	+	–	–
Acer sp.	–	–	–	+	–	Gramineae gen. indet.	–	–	+	+	+
Ilex colchica A. Pojark.	+	–	+	+	–	*Typha minima*	–	–	–	+	–
Euonymus sp.	–	–	+	–	–	*Cyperites braunianus* Heer.*	+	–	–	–	–
Staphyllea sp.	–	–	–	+	–	Cyperaceae gen. indet.	–	–	–	+	+

* Tertiary relics extinct or missing in the recent flora of western Georgia.

dark coniferous forests and never was prevailing in mixed forests of the lower mountain zone usually dominated by *Carpinus* and frequently by *Castanea* groupings. Similar forests of Abkhazia are described by Dolukhanov (1938). Pure beech forests could be characteristic of the middle mountain zone. Beech trees likely were more widespread in northern Kolkhida than in its southern territory. *Tilia* and *Fraxinus* were accessory amid principal taxa.

Hemihygrophilic mixed forests that consisted of *Castanea*; *Fagus*; *Acer pseudoplatanus* L.; *A. polymorphum pliocenicum* Sap.; and undergrowth of *Buxus*, *Hedera*, *Ilex*, and other taxa populated the mountain canyons. Areas of plant formations of the Mediterranean maquis type periodically increased and decreased. The dark coniferous forests of the high mountain zone migrated downward three times during Stage II approaching the piedmont zone. In the periods of humidity decrease, they were enriched in spruce species, whereas elm trees grew in abundance every time when the competition ability of principal dominant species decreased in response to a noticeable cooling, excessive aridness, or elevated humidity. Diversity of plant assemblages and their distribution patterns directly reflected variations in heat supply and humidity.

In the climate evolution, we distinguished four warming periods separated by three epochs of cooling. It is evident that warm climatic phases display rather regular changes in their duration and intensity: each of them was longer and warmer than the preceding one. It is more difficult to judge the character of cooling phases, because the respective intervals of sections are often composed of coarse-grained deposits lacking phytological remains. However, it also clear that the cooling period in Stage II was most intense, and another one in the next stage was considerably less significant. We consider the last stage as corresponding to the Eltingen warming fixed in the climatostratigraphic scheme of the Black Sea region (Arkhangel'skii and Strakhov, 1938; Andrusov, 1965; Zubakov *et al.*, 1982). In addition, we distinguished a horizon at the depth of 44.3 m, where the pine pollen content in SPA is as high as 30% (for comparison, SPA of the first Würm stadials contain 40–54% of the pine pollen). Such a significant indication of cooling allows us to consider this event as corresponding to the Geroevskoe-2 cooling phase. Subsequently, the lowermost beds of the studied section (the first of our stages) may be interpreted as deposits that have been accumulated in paleogeographic environments of the Zavetnino warming phase (Zubakov *et al.*, 1982; Zubakov, 1992).

During Stage I equivalent to warming phase of the Zavetnino time (190 ka ago), the climate was more warm and somewhat less humid than at present. Air temperature of that time was by 1.2–2°C higher in general (Zubakov, 1986). In the lower course of the Enisei River, water and air temperatures were higher by 1–2°C and 2–3°C respectively (Troitskii, 1979).

Stage II, an equivalent of the cooling phase Geroevskoe-2 (165 ka ago), corresponds to the colder and more humid climatic period as compared with the recent time. Dark coniferous forests then descended quite low, maybe very close to the piedmont zone. This was also a period of water outflow from the Caspian Sea through the Manych River valley (Goretskii, 1982; Popov, 1983) and from the Black Sea through the Bosporus to the Mediterranean Sea (Zubakov, 1986).

Stage III, an analogue of the Eltingen warming phase (127–115 ka ago), is especially interesting from the viewpoint of predicting the future climatic conditions. It corresponds to the time of the Mikulino Interglacial in the Russian plain, Eemian–Sangamonian event in northwestern Europe and America, and Kazantsevo warming in West Siberia. Numerous data considered by Zubakov (1986, 1992) suggest that the early Karangat warming spread over vast areas of the northern hemisphere, Georgia included.

We should note that our results are concordant with other data about the evolution of the early Karangat flora in Georgia. For instance, the upward migration of vegetation zones is recorded in flora from travertine beds in the Dolra River valley (one of tributaries of the Inguri River), where, at the altitude of 2200 m, it contains remains of *Ulmus foliaceae* Gilib. and *Ostrya carpinifolia* Scop. unknown now at this level. The vegetation composition (Table 2) shows that the upper marginal forest zone was formerly located higher than now (Ratiani, 1964). Considering other Georgian floras compatible with the climatic optimum, we tried to evaluate their position in our scheme of the Riss–Würm climatostratigraphic subdivisions (Table 1).

Flora from the Potara Poti Well 3 (Mamatsashvili, 1975) distinctly reflects two epochs of very warm and relatively arid climate similar to those in our substages 1 (Phase 1b) and 3 (Phase 1b) and 3 (Phase 3b). Two other events marked by the unusually high abundance of beech pollen clearly correspond to past periods of a cooler climate, when the beech forest zone migrated downward. These events are correlative with our substages 2 and 4 (Phase 4a). Flora from Kobuleti localities (Shatilova and Badzoshvili, 1966) is again indicative of a warm and dry climate. Its general composition and proportions of species, except the higher content of *Fagus* specimens, are similar to these parameters of our Substage 3 (Phase 3b).

The Mt. Sukhumi flora (Kolakovskii, 1959) is an indicator of a rather long and arid climatic period analogous to the Karangat Phase 1b. This flora is enriched in xerophilous species (14 of 43 in total), such as *Quercus cerris* L., *Zelkova carpinifolia* (Pall.) Dipp., and some others. In addition, it includes a series of taxa characteristic of the Mediterranean maquis. In contrast, flora of the Gumista River localities (Ratiani, 1960, 1970) is depleted in xerophilous elements (3 species only), but contains abundant macroremains of *Fagus*, *Alnus hoernesi* Stur, *Parrotia pristina* Ett., and *Salix*

alba L. (Table 1) characteristic of a warm and humid climatic epoch comparable with our phases 4a and 4b.

Thus, the early Karangat evolution of climate and vegetation in Abkhazia is characterized by the principal trends described above. The general list of species (about 160 in total), which are described at the moment, is presented in Table 2. One fifth of them belongs to species extinct now or growing far away from the region of our study. The later Karangat and Neoeuxinian floras are considerably less diverse and consist almost entirely of taxa characteristic of the modern vegetation in Abkhazia.

CONCLUSION

Palynological characteristics of the lower Karangat deposits show that the vegetation distribution in this period was inconstant. Not only dark coniferous formations, but also broad-leaved forests confined to lower and intermediate mountain zones, periodically migrated in space (Fig. 2).

The inferred changes in vegetation and climate reflect the three main stages in their evolution. The dark coniferous forests migrated three times downward and approached the piedmont zone most closely in the second stage. The latter was the "coldest" in the considered time interval (Fig. 2*b*). In contrast, the third stage was the warmest and consisted of four substages and several climatic phases. This Riss–Würm climatic optimum is marked by the predominance of *Carpinus* and *Castanea* forms. All exotic species, particularly, Taxodiaceae, were noticeable components of flora at this period. At the same time, the dark coniferous forests were displaced high in the mountains, and the geographic range of this zone was considerably reduced (Fig. 2*c*).

Throughout the Riss–Würm or early Karangat, the lowland landscapes were occupied mainly by hornbeam–alder and, locally, by pure alder forests, the latter consisting of *Alnus hoernesi* and *Alnus barbata*. Near rivers, the forests included subordinated *Pterocarya* and abundant *Salix alba*. Pine forests were also widespread. Groupings of Taxodiaceae, *Cedrus*, *Carya*, and *Tsuga* were under gradual decline. Judging from macroremains, *Parrotia pristina* was one of important elements of vegetation.

The Riss–Würm evolution of climate in Abkhazia included four warming stages and three epochs of cooling between them. We revealed a distinct trend in the evolution of the warming stages: each of them was longer and warmer than the previous stage. In addition to this general trend toward the more arid and warmer climate, there were subordinate climatic fluctuations, when conditions turned out to be either colder and more humid, or warmer and humid, and even similar to modern environments. The relatively dry conditions were characteristic of the middle first substage and of the end of third substage of the last stage. The second stage apparently was most humid.

Data on the pollen content of aquatic and riparian plants from the seashore zone reveal the evolutionary trend of lagoonal lakes (Balabanov and Kvavadze, 1985). Pollen of riparian plants is most abundant in the Karangat deposits of the first stage. In sediments of the two next stages, this pollen appears periodically. Later, during the Würm and Holocene, lagoonal lakes apparently disappeared from the region in question, because the respective deposits do not contain pollen of aquatic plants. These data suggest that exactly in the early Karangat the water level in the Black Sea was higher than in other periods of the late Pleistocene.

ACKNOWLEDGMENTS

This work was partially supported by the International Science Foundation, grant MXJ000.

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